GEOTHERMAL FLUID GENESIS IN THE GREAT BASIN

FINAL REPORT

FOR THE PERIOD

1 AUGUST 1988 - 1 MAY 1990

WORK PERFORMED UNDER CONTRACT

DE-FG07-88ID12784

DIVISION OF EARTH SCIENCES
ENVIRONMENTAL RESEARCH CENTER
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DOE/ID/12784--1

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U.S. DEPARTMENT OF ENERGY
IDAHO OPERATION OFFICE
UNDER CONTRACT DE-FG07-88ID12784

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ACKNOWLEDGEMENTS

After eighteen gruelling months of scientific investigative research, the authors would like to acknowledge the people that assisted with the successful completion of this project. By all accounts, the following individuals deserve more than just mere praise for their unselfish willingness to be abused, offended, and outraged by the undisguised gall of the principal investigators. But that's all they are going to get.

Many thanks to the entire Nevada geothermal community, especially Stu Johnson, Von Maynard and Fred Manuel of Chevron Resources, Dick Benoit of Oxbow Geothermal, Pamela and Mike Lattin of Elko Heat Company, Francis and Gordon McKay of Warren Estates, Al Hansen and Mark Keesee of Geothermal Food Processors, and the Elko County School District. Thanks to Paul Hirtz, Jeane Stewart, and Cliff Buck of Thermochem Inc. for all their help with sampling.

Larry Benson and Robert Thompson of the U.S. Geological Survey supplied us with some data and ideas early in the project. We thank Jim (Bo) Obar and the good people at Death Valley National Monument for their help with processing our sampling permit in record time. Alan Welch of the U.S. Geological Survey gets two stars, one for loaning us the carbonate precipitator and the other for reviewing this heresy. Thanks to Peter Wigand of the Desert Research Institute, who supplied us with freshly culled macrofossils from selected packrat middens. Merci to Austin Long, who arranged for cellulose separation and deuterium analysis. Thanks to Linda Blair of the Environmental Research Center, UNLV, for her archaeological contribution.

A big hand for the review team, Dennis Trexler of the Division of Earth Sciences, UNLV, Howard Ross of the University of Utah Research Institute (who was also the contract monitor), Robert Mariner of the U.S. Geological Survey, Larry Garside of the Nevada Bureau of Mines and Geology, and Marshall Reed of the U.S. Department of Energy in Washington, D.C. The authors also appreciate the support and patience of the U.S. DOE, Idaho Operations Office, especially Ken Taylor, Ken Osborne, and Trudy Thorne.

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ABSTRACT

Early theories concerning geothermal recharge in the Great Basin implied recharge was by recent precipitation. Physical, chemical, and isotopic differences between thermal and non-thermal fluids and global paleoclimatic indicators suggest that recharge occurred during the late Pleistocene.

Polar region isotopic studies demonstrate that a depletion in stable light-isotopes of precipitation existed during the late Pleistocene due to the colder, wetter climate. Isotopic analysis of calcite veins and packrat midden megafossils confirm the depletion event occurred in the Great Basin.

Isotopic analysis of non-thermal springs is utilized as a proxy for local recent precipitation. Contoured plots of deuterium concentrations from non-thermal and thermal water show a regional, systematic variation. Subtracting contoured plots of non-thermal water from plots of thermal water reveals that thermal waters on a regional scale are generally isotopically more depleted. Isolated areas where thermal water is more enriched than non-thermal water correspond to locations of pluvial Lakes Lahontan and Bonneville, suggesting isotopically enriched lake water contributed to fluid recharge. These anomalous waters also contain high concentrations of sodium chloride, boron, and other dissolved species suggestive of evaporative enrichment.

Carbon-age date and isotopic data from Great Basin thermal waters correlate with the polar paleoclimate studies. Recharge occurred along range bounding faults.

CHAPTER 1. INTRODUCTION

Geothermal energy resources are widespread in the Great Basin and presently account for nearly 200 megawatts of electrical power production. In Nevada, there are more than 900 geothermal springs and wells. In addition to electric power production, geothermal energy is currently used for space heating and industrial processes throughout Nevada, Utah, California, and Oregon. Virtually all of the known geothermal resources in the Great Basin are classified as liquid dominated. That is water, rather than steam, is the primary fluid in the reservoir. The source of that fluid has long been the focus of scientific investigations in the disciplines of geology, hydrology, and reservoir engineering. Due to the recent widespread development of geothermal resources, government agencies involved in regulation of the geothermal industry have a vested interest in all aspects of geothermal energy, including models of geothermal fluid recharge.

Geothermal fluids throughout the Great Basin are significantly different, physically chemically and isotopically, when compared to non-thermal waters. The most obvious physical difference is the elevated temperature, which is generally attributed to circulation within a zone of high heat flow. Chemical differences include anomalous trace constituents and high concentrations of total dissolved solids. These may be accounted for by high-temperature interaction with carbonate and volcanic rocks and marine sediments.

Another difference between thermal and non-thermal fluids in the Great Basin is in the ratio of stable isotopes of hydrogen and

oxygen. Geothermal fluids are typically isotopically more depleted (lighter) than non-thermal fluids issuing at the same elevation. Isotopic differences in natural waters result from variations in temperature, elevation, and other physical factors.

Initial hypotheses concerning geothermal recharge demonstrated the similarity between the isotope content of geothermal fluids and precipitation falling on nearby mountain highlands. According to this theory, run-off from precipitation falling in the mountains infiltrates downward through fractures to a depth of three to five kilometers, is heated, and returns to the surface. The theory accounts for isotopic differences, but is self-limiting because it requires the close association of high-elevation recharge areas that can accumulate sufficient volumes of fluid to recharge large geothermal systems. As will be demonstrated, many geothermal areas in the Great Basin are not located sufficiently close to elevated recharge areas to depend solely on this mechanism.

An alternative explanation proposes that geothermal fluids may have been recharged during the Pleistocene, when mean temperatures were decidedly cooler and precipitation amounts greater. This hypothesis also accounts for the isotopic differences and is not self-limiting; it does not rely on specific topography. It will be shown that global-scale paleoclimatological data indicate a major isotope depletion occurred late in the Pleistocene Epoch, a period of time when glaciers covered large portions of North America and Europe.

The purpose of this investigation is to develop a conceptual model for paleo-geothermal fluid recharge in the Great Basin. The basis for this model is the isotopic analysis of fluid samples collected from deep, high temperature geothermal production wells and paleoclimatological proxy data, such as ice-core and packrat midden data, used to de-limit the temporal framework of the model. Isotopes from geothermal fluids throughout the Great Basin will be shown to vary systematically in a manner similar to present-day non-thermal fluids. In addition, radiocarbon age dating techniques have been applied to thermal fluids and to paleoclimate proxies. The model will show that stable isotopes from precipitation in the Great Basin during the last 40,000 years also demonstrates an isotopic depletion during the late Pleistocene. Further, it will demonstrate that the high-stands of the pluvial lakes that occupied basins in Nevada and Utah, are intimately associated with the isotopic depletion and affected the mechanism by which geothermal fluid recharge occurred.

CHAPTER 2. PHYSICAL SETTING OF THE GREAT BASIN

The Great Basin is a topographically elevated and relatively dry region that is characterized by north and northeast trending mountains and valleys. It is geologically complex, and has a relatively high heat flow. This combination of factors has produced the present distribution of geothermal systems.

LOCATION AND GEOGRAPHY

The Great Basin of the western United States (Figure 2.1) is a physiographic province characterized by north to northeast trending mountains and valleys, and defined by a regional internal drainage system (Fenneman, 1931). Elevations range from 1,300 to 1,600 meters (m) in the valleys, to 3,000 to 3,600 m for the highest peaks. The Great Basin is intimately associated, and often confused, with the Basin and Range Province, a regionally extensive geologic province of block faulting which forms the aforementioned north to northeast trending mountains and valleys. The Basin and Range Province is larger than the Great Basin, and includes large parts of Arizona and New Mexico.

Rainfall varies greatly throughout the region, controlled by latitude, longitude, and elevation. The higher ranges receive 20 to 40 centimeters (cm) of precipitation per year, usually between November and March. Precipitation at lower elevations ranges from 0 to 10 cm per year. Valley playas are generally dry most of the year, but are the site of ephemeral lakes during periods of high runoff or prolonged precipitation.

Figure 2.1: Location map for the western United States, and the Great Basin physiographic province as defined by the dashed line.

GEOLOGY

The Great Basin is geologically complex, and a comprehensive description is beyond the scope of this report. However, a general description of the geological history and principal features will serve as a framework for the ensuing discussions.

Rock units in the Great Basin range in age from Precambrian to recent. The oldest rocks, ranging in age from Precambrian through middle Paleozoic, occur in the southern and eastern portion of the province. They are the product of extensive clastic and carbonate deposition in shallow seas, with the estimated thickness of the sedimentary assemblages in eastern Nevada and western Utah exceeding 6,000 m. Paleozoic and Mesozoic rocks in western Nevada are largely deep sea shales and siliceous siltstones that were thrust over the eastern assemblage carbonates in a series of orogenies that began in the Devonian Period, 370 million years ago (mya), and continued through the middle of the Mesozoic Era, 170 mya. Much of the extreme western Great Basin is underlain by granitic plutons of Mesozoic age, which are best exposed in the Sierra Nevada Mountains of western Nevada and eastern California.

Throughout the Great Basin, the valleys are filled with sediments derived from the local ranges. Sediment size varies from very coarse fan-glomerates near the ranges, to extremely fine silts, clays, and evaporites in the central valley playas. Valley sediment accumulations may be as thick as three kilometers (km).

Widespread volcanism began approximately 43 mya and resulted in the formation of siliceous caldera complexes. Volcanic activity initiated in western Utah and eastern Nevada, spreading west and south across the region. Approximately six mya, the volcanism changed from bimodal (siliceous and basaltic), to principally basaltic. The most recent episode of volcanism was basalt-basaltic andesite lava flows and cinder cones in west central Nevada, less than 100,000 years ago.

Widespread faulting throughout the region was also initiated approximately 43 mya, and is thought to be associated with the volcanic activity. Basin and Range faulting, which produced the present-day topography of elongated mountains and valleys, is the result of regional extensional tectonics that began approximately 13 mya. This block faulting style produces sub-parallel mountain ranges that are, on average, 10 to 20 km wide and measure 25 to 35 km from crest to crest. Several models have been developed to explain this structural style (Figure 2.2). A generalized map of the late Cenozoic extensional faults (Figure 2.3) in the Great Basin shows their wide distribution and north-northeast trend. Range bounding faults are responsible for the widespread earthquake activity in the Great Basin, and are also believed to be the principal conduits for convecting geothermal fluids.

THERMAL CHARACTERISTICS

In addition to a complex geologic history, the Great Basin is located within an area of elevated heat flow. Heat flow is defined

as the product of the thermal conductivity of a rock type and the measured thermal gradient. Blackwell (1983) argued that the high observed heat flow in the Great Basin is a product of the superposition of the regional effects of tectonic extension, surface volcanism and subsurface igneous intrusion, modified by the local conductive effects of thermal refraction, variations in radiogenic heat production, erosion and sedimentation. A conceptual heat flow model was proposed that included a heat source, emplaced at a depth of 10 to 20 km, with a temperature of $1,350^{\circ}\text{C}$. Any disturbances in observed heat flow at the surface are related to deep circulation of groundwater. An extended, highly permeable upper crust and the relatively high geothermal gradient of the Great Basin controls the distribution of heat flow and geothermal systems.

Figure 2.2: Models of Basin and Range structure, after Stewart (1983).

GEOHERMAL RESOURCES IN THE GREAT BASIN

Although geothermal resources, in general, have long been the focus of interest and curiosity, serious scientific investigations were not initiated in the United States until the mid 1970's. Earlier in the decade, Congress passed the Geothermal Steam Act, which established the framework for exploration and development of steam and hot water that was to be found on public lands in the United States.

To determine the value and effectiveness of developing known geothermal resources, a series of assessment programs was initiated

Figure 2.3: Distribution of late Cenozoic extensional faults in the Great Basin, from Stewart (1983).

by various state and federal agencies. One of the first of these was a nationwide assessment of geothermal resources (White and Williams, 1975). The document provided an analysis of steam and liquid dominated systems, igneous related resources, and hot dry rock (HDR). In Nevada, the authors identified 18 separate liquid dominated systems and classified them largely on the basis of geochemical parameters and surface manifestations. A companion section described regional heat flow; anomalous values, as high as 3 heat flow units (HFU), were determined for an area in northern Nevada near the town of Battle Mountain. In 1971, Sass et al. first described the anomalously high temperatures in shallow wells and coined the term "Battle Mountain Heat Flow High".

Garside (1974) documented the initial exploration efforts in Nevada, and identified approximately 300 thermal springs and wells throughout the state. Included in the report were thirteen areas that the U.S. Department of the Interior had previously classified as Known Geothermal Resource Areas (KGRA's). These are large areas of mostly public land that the Federal Government believed had the potential for development. These areas were subsequently offered for lease according to the provisions of the Geothermal Steam Act.

Many of the earliest assessment programs incorporated elements of geological reconnaissance, geophysical surveys, geochemical evaluations, and drilling. Mariner et al. (1974) tabulated the available chemical data for the largest hot springs systems in northern Nevada and provided estimates, on the basis of chemical geothermometers, of subsurface temperatures. In 1975, Mariner et

al. completed a report on trace elements and gas compositions for hot springs in Nevada and Oregon, and initiated the compilation of a geothermal data base for Nevada. Mariner et al. (1976) appended the data base with chemical information from hot springs located in three KGRA's. In addition to geochemical data, the authors also used stable light-isotopes to characterize the geothermal systems, noting a similar "oxygen isotope shift" for hot springs in the region.

Hose and Taylor (1974) proposed a model for Basin and Range geothermal systems in which deeply penetrating fault zones, in a region of elevated thermal gradients ranging from 40°C to 60°C per km depth, provided pathways for the downward passage of meteoric water and the upward migration of thermal fluids. Wollenberg et al. (1975) completed a geothermal resource assessment of northern Nevada, based upon an integrated program of geology, geophysics, and geochemistry, of four high temperature thermal spring systems: Beowawe, Buffalo Valley, Leach Hot Springs, and Kyle Hot Springs. Their conclusions concurred with the concept of a close association between thermal springs and fault zones, and contributed to the development of a conceptual model of geothermal systems. The model suggested that recharge was derived from meteoric water saturating alluvial fans. The water percolated downward to a depth of four to five kilometers, reaching temperatures ranging from 150°C to 200°C. Range-bounding fault zones were suggested as the permeable passageway for the upwardly migrating hot water.

As the volume of data for geothermal systems grew, and in the wake of the "energy crisis" in the mid to late 1970's, efforts were

accelerated to develop geothermal energy in the United States. Muffler (1978) completed another assessment of geothermal resources in the United States and classified hydrothermal systems on the basis of chemical geothermometers. In Nevada, he identified 17 systems with estimated subsurface temperatures in excess of 150°C, 24 systems between 90°C and 150°C, and 28 systems less than 90°C. Virtually all of the systems were believed to result from deep circulation along fault zones.

Following the regional assessment phase, efforts were focused on "site specific" investigation of hot spring systems. These investigations included summaries of existing data, and attempted to provide conceptual models of the individual hydrothermal systems on the basis of combined geological, geophysical, and geochemical data. Many studies concentrated on Long Valley Caldera, eastern California (Bailey, 1974, 1982; Dimet et al., 1980; Fournier et al., 1979; and Lachenbruch and Sass, 1977). An extensive review of the Long Valley studies and system is provided by Higgins et al. (1985). Additional site specific studies were completed in other parts of eastern California (Levias et al., 1981; Levias and Bacon, 1982; and Higgins et al. 1983), and serve as the foundation for the data base in that portion of the study area.

Garside and Schilling (1979) provided an inventory of Nevada's geothermal resources that included chemical and isotopic data, in addition to development histories of selected geothermal areas. Trexler et al. (1979) completed an assessment of Nevada's low- to moderate-temperature thermal resources that identified more than

320 geothermal springs and wells and classified them on the basis of temperature and fluid chemical composition. Following this, a series of assessment programs were completed in various portions of Nevada including Caliente, (Trexler et al., 1980a), Carson-Eagle Valley and Big Smoky Valley, (Trexler et al., 1980b), Hawthorne, Paradise Valley, and the Carson Sink, (Trexler et al., 1981), and Pumphernickel Valley, Carlin, and Moana (Trexler et al., 1982). These studies consisted of a series of synoptic surveys, including geology, geophysics, and geochemistry. In most cases, the near surface geothermal systems were found to be fault-controlled and both chemically and isotopically distinct from non-thermal fluids.

A significant effort was focused on the development of high temperature geothermal resources in Utah and Nevada in the late 1970's to early 1980's. This project was known as the Industry Coupled Drilling Program and was jointly administered by the U.S. Department of Energy and the geothermal industry. In Nevada, large scale drilling programs were completed at ten sites: Beowawe, Dixie Valley, Stillwater, Humboldt House, Desert Peak, Soda Lake, Colado, Leach Hot Springs, McCoy, and Tuscarora. In Utah, the drilling program included Cove Fort and Roosevelt Hot Springs. Both of the Utah sites and five of the ten Nevada Sites presently support a geothermal power plant. New chemical and isotopic data from the Beowawe, Dixie Valley, Desert Peak and Soda Lake plants are used in this report.

Benoit et al. (1982) completed a case history of the Desert Peak geothermal area that documented the development of an exploration strategy designed to identify "blind" geothermal resources;

geothermal systems with no surface manifestations. The report includes detailed chemical and temperature data for the field.

The physical and chemical characteristics of high-temperature geothermal systems in the northern Great Basin were documented by various authors (GRC, 1983). The relationship between observed anomalous heat flow and the development of energy and mineral resources was illustrated.

Edmiston and Benoit (1985) reviewed the characteristics of six geothermal systems with temperatures between 150° and 200°C. They concluded that most of studied systems were associated with hot springs that discharge sodium chloride type fluids in northwest Nevada, or with major centers of Quaternary volcanism along the western border of the Great Basin. Two exceptions are the Cove Fort and Roosevelt fields in southeastern Utah. They also noted that the silica-based chemical geothermometers provided the most favorable data for high-temperature exploration programs.

Thermal fluids with temperatures in excess of 35°C in western and central Utah were documented by Goode (1978). The report lists chemical and/or isotopic data for more than 1,500 thermal wells and springs. Mabey and Budding (1987) described seven high-temperature (>90°C) geothermal systems in the Sevier geothermal area of southwestern Utah. Although no isotopic data are used, the authors classified thermal fluids throughout the region using a geochemical signature based on common ion chemistry.

Chemical and isotopic data were used together to delineate a portion of the Roosevelt geothermal system as an area to supply make-up water for a hot-dry-rock energy extraction system (Vuataz and Goff, 1987). The isotopic data on the thermal fluids show a slight deuterium depletion and oxygen shift, relative to the non-thermal fluids in the area.

In summary, geothermal systems in the Great Basin have an extensive history of exploration and subsequent development. The presently available data base consists of an integrated package of geological, geophysical, and geochemical surveys, supplemented by subsurface temperature and lithologic logs. Most of the geothermal power plant production wells produce a combination of brine and steam, and were developed in the vicinity of known high-temperature geothermal springs. The location of other geothermal developments are intimately associated with Quaternary age volcanism.

CHAPTER 3. GEOCHEMICAL DATA BASE

DATA ACQUISITION

Geochemical and isotope data from thermal and non-thermal springs and wells throughout the Great Basin were assembled over a period of approximately twelve months. A computer search utilizing the GEOREF, CHEM ABSTRACTS and POLLUTION ABSTRACTS data bases was initiated using the following key word combinations: "Nevada" or "Utah" or "Great Basin;" and "stable light-isotopes" or deuterium" or "oxygen-18" or "tritium;" and "geothermal" or "water." This search was augmented by investigations of publication lists from:

Utah Geological and Mineralogical Survey,
Nevada Bureau of Mines and Geology,
California Division of Oil and Gas,
Division of Earth Sciences, University of Nevada, Las Vegas,
Desert Research Institute, University of Nevada,
Theses and Dissertations from various western Universities.

Chemical data were also obtained from the Division of Water Resources and the Division of Environmental Protection, Department of Conservation and Natural Resources, State of Nevada. The Department maintains records on geothermal fluid injection wells, and part of the mission is to maintain a data base of geochemistry. Information from operating geothermal power plants, industrial process plants, and space heating districts was obtained.

One unanticipated source of information was the data set established during site characterization of the proposed high level nuclear waste repository at Yucca Mountain, Nye County, Nevada. Interdisciplinary studies under way include groundwater hydrology, paleoclimatology, and geochemical surveys.

During the data base acquisition period, several data void areas were identified. Isotopic data from the southern and western portions of the study area were lacking. In addition, there was a lack of recent isotopic data from the deep, geothermal power plant production wells. The fluid sampling program was designed to optimize the data set by focusing principally on identified data gaps. Contacts with geothermal power plant operators, established prior to the start of the contract, were re-established to obtain archived historical data and samples of geothermal fluids from deep reservoirs.

FIELD SAMPLING PROGRAM

As discussed above, the purpose of the field sampling program was to acquire data from "void" areas, and to acquire new data from deep production wells at geothermal power plants. In addition, samples from hot springs with an established history of geochemical investigations were obtained in order to identify compositional variations over time.

A total of 42 fluid samples were collected in the course of the investigation, including 42 for stable light-isotope (deuterium and oxygen-18) analysis, 38 for bulk chemical analyses, 16 for

tritium analysis, and 12 for carbon-14 age dating. Figure 3.1 shows the locations of the fluid sampling sites and Table 3.1 lists the data associated with each sample. Field sampling techniques and analytical laboratory methods are listed in Appendix 1.

Brine and steam samples were collected from the Desert Peak Power Plant and Beowawe Power Plant, operated by Chevron Resources Company. Brine samples were also obtained from the Soda Lake Power Plant, operated jointly by Chervron and ORMAT Energy Systems. Samples were also obtained from the Wabuska Power Plant, operated by TAD'S Enterprises, the Elko County Municipal Heating District,

Figure 3.1: Location of fluid samples taken for isotopic and chemical analysis during this study. Refer to GF- numbers in Table 3.1 for analyses.

Table 3.1 - Current study analyses of stable light-isotopes and geochemistry. See Page 21 for key to basin locations (LOC).

GF #	SAMPLE NAME	REF	LOC	TYPE	LONG	LAT	ELEV meters	DEUT 0/00	OX18 0/00	TEMP °C	DATE MO YR
1	Bradys Hot Springs	GFG	BDY	WELL	119.012	39.787	1253	-127	-14.2	141	1 89
2	Great Boiling Spring	GFG	BRD	SPRG	119.365	40.661	1207	-103	-10.9	91	1 89
3	Patua Hot Spring	GFG	CSK	SPRG	119.113	39.597	1237	-121	-12.4	86	1 89
4	Steamboat/Ormat	GFG	SMA	WELL	119.715	39.395	1433	-121	-12.4	113	1 89
5	Wabuska Tads #2	GFG	MSV	WELL	119.152	39.161	1311	-115	-15.8	107	1 89
6	Warren Estates #1	GFG	SMA	WELL	119.825	39.481	1402	-126	-15.9	88	1 89
7	Soda Lake 33-14	GFG	CSK	WELL	118.859	39.564	1213	-105	-10.8	183	2 89
8	Desert Peak 67-21	GFG	DPK	WELL	118.951	39.754	1340	-115	-12.3	158	2 89
9	Desert Peak 86-21	GFG	DPK	WELL	118.946	39.758	1366	-114	-12.1	159	2 89
10	Beowawe Ginn 1-13	GFG	BEO	WELL	116.616	40.554	1506	-132	-16.2	151	2 89
11	Beowawe Ginn 2-13	GFG	BEO	WELL	116.617	40.555	1506	-133	-15.9	153	2 89
12	Lee Hot Springs	GFG	FAL	SPRG	118.722	39.209	1222	-126	-13.4	93	2 89
13	Alkali Hot Spring	GFG	CLV	SPRG	117.335	37.825	1529	8	8	52	2 89
14	Hicks (Bailey) HS	GFG	AMG	SPRG	116.721	36.805	1097	-116	-14.5	40	2 89
15	Grapevine Spring	GFG	DVY	SPRG	117.384	37.024	756	-112	-13.9	32	2 89
16	Nevares Spring	GFG	DVY	SPRG	116.791	36.512	293	-105	-13.4	31	2 89
17	Tecopa Hot Spring	GFG	DVY	SPRG	116.231	35.872	427	-99	-12.6	42	2 89
18	Rogers Spring	GFG	MRV	SPRG	114.443	36.377	485	-95	-13.7	31	2 89
19	Ash Spring	GFG	WRV	SPRG	115.195	37.465	1100	-111	-14.2	30	2 89
20	Warm Spring	GFG	HCV	SPRG	116.377	38.187	1682	-118	-14.2	61	2 89
21	Golconda Hot Sp	GFG	PDV	SPRG	117.491	40.955	1341	-132	-15.8	58	2 89
23	Elko Jr Hi	GFG	ELK	WELL	115.761	40.839	1554	-151	-17.9	82	2 89
24	Elko Heat Company	GFG	ELK	WELL	115.775	40.825	1541	-149	-18.1	80	2 89
25	Bruffeys Hot Sp	GFG	PNE	SPRG	116.067	40.067	1783	-133	-16.3	62	2 89
26	Williams Hot Sp	GFG	WRV	SPRG	115.231	38.953	1931	-121	-15.6	53	2 89
27	Morman Spring	GFG	WRV	SPRG	115.142	38.763	1609	-125	-16.1	38	2 89
28	Chimney Hot Sp	GFG	RRV	SPRG	115.788	38.467	1465	-124	-15.9	66	2 89
29	Hot Creek (Upper)	GFG	HCV	SPRG	116.361	38.521	1725	-115	-14.2	82	2 89
30	Darroughs Hot Spring	GFG	BSV	WELL	117.183	38.821	1695	-125	-15.7	80	2 89
31	Surprise Valley hs	GFG	SPV	SPRG	119.978	41.166	1388	-127	-15.4	47	3 89
32	Soda Lake 33-14	GFG	CSK	WELL	118.859	39.564	1213	-104	-11.2	183	5 89
33	Desert Peak 67-21	GFG	DPK	WELL	118.951	39.754	1340	-118	-12.1	163	5 89
34	Beowawe 1-13	GFG	BEO	WELL	116.616	40.554	1506	-131	-15.9	150	5 89
35	Wabuska Tads #2	GFG	MSV	WELL	119.152	39.161	1311	8	-16.2	107	5 89
36	Warm Spring	GFG	HCV	SPRG	116.377	38.187	1682	-119	-14.8	61	5 89
37	Elko Jr Hi	GFG	ELK	WELL	115.761	40.839	1554	-150	-18.6	80	5 89
38	Elko Heat Company	GFG	ELK	WELL	115.775	40.825	1541	-147	-18.4	81	5 89
39	Golconda Hot Spring	GFG	PDV	SPRG	117.491	40.955	1341	-132	-16.3	63	5 89
44	Soda Lake 33-14	GFG	CSK	WELL	118.859	39.564	1213	-106	-11.2	185	9 89
46	Beowawe Ginn 2-13	GFG	BEO	WELL	116.617	40.555	1506	-131	-16.1	154	9 89
47	Steamboat/Ormat	GFG	SMA	WELL	119.715	39.395	1433	-119	-13.0	113	9 89
48	Wabuska Tads #2	GFG	MSV	WELL	119.152	39.161	1311	-133	-16.4	107	9 89

Sample locations are shown in Figure 3.1, except for samples GF-32 to 48 which are duplicates. Sample GF-22 was not submitted for analysis. B - Broken in shipment. pH's are laboratory measurements at room temperature.

Table 3.1, continued

Current Study Data Set

GF	pH	Ca	SO ₄	Na	K	SiO ₂	CO ₃ + HCO ₃	Cl	Mg	Li	F	B	SPEC COND	SUM TDS	MEAS TDS
#				(All)	Analyses		Reported	In	PPM)				mmhos		
1	7.2	35	323	694	53	210	112	872	0.2	1.5	5.5	4.4		2311	2120
2	7.3	73	416	1488	113	169	97	2180	1.4	1.7	4.4	8.2		4552	4152
3	7.6	52	405	656	52	198	93	829	0.6	1.7	4.7	6.1		2298	2100
4	6.3	15	120	611	58	278	369	790	0.3	6.9	2.0	41.8		2292	2056
5	7.5	35	549	292	14	111	73	47	0.0	0.3	6.8	1.0		1129	1038
6	8.1	27	528	277	8	126	93	55	0.0	0.2	4.1	2.0		1120	950
7	5.7	109	48	2000	232	284	108	3400	0.4	3.8	1.0	13.4	10000	6200	6140
8	8.1	96	77	2190	235	303	38	3550	0.2	3.0	4.6	15.1	12000	6512	7320
9	8.0	87	70	2230	249	319	36	3740	0.2	3.0	4.3	15.6	12000	6754	7570
10	9.7	0	74	169	21	335	213	41	0.2	1.0	9.6	0.9	850	865	887
11	9.8	0	75	171	21	308	163	42	0.2	1.0	8.8	0.9	850	791	824
12	7.3	40	465	458	26	164	114	381	0.6	0.7	7.0	2.2		1659	1601
13	7.0	45	504	320	22	55	326	62	3.1	1.7	6.7	1.1		1347	1116
14	7.7	17	139	168	7	65	228	54	0.5	0.3	5.2	0.3		684	542
15	7.8	41	138	183	15	44	453	80	19.1	0.2	2.7	1.1		977	680
16	7.6	44	179	148	11	26	360	44	21.2	1.2	2.6	0.9		838	600
17	8.4	4	569	798	16	89	713	439	1.7	0.2	2.5	8.8		2641	2148
18	7.2	443	1740	323	21	20	163	338	140.4	0.6	1.4	0.9		3191	3064
19	7.6	47	37	28	6	29	257	14	19.0	0.0	0.5	0.1		438	272
20	6.8	70	98	193	23	48	704	36	22.0	0.6	3.0	0.4		1198	798
21	7.3	33	59	147	22	60	449	22	7.4	0.5	1.6	1.4		803	580
23	7.0	61	77	114	37	73	496	15	11.0	0.3	1.5	1.0		887	600
24	6.6	63	70	110	35	65	493	15	12.3	0.3	2.0	0.9		867	582
25	7.0	47	37	39	8	55	279	13	17.1	0.1	0.7	0.3		496	326
26	9.6	2	15	63	0	59	60	11	0.2	0.0	13.8	0.2		224	180
27	7.5	62	46	27	5	27	296	11	19.8	0.1	1.3	0.1		495	328
28	6.8	68	48	67	15	43	412	11	14.0	0.2	1.8	0.3		680	448
29	7.0	24	58	208	13	162	524	43	4.0	1.0	8.8	0.6		1046	745
30	9.2	1	47	104	2	104	126	13	0.0	0.3	14.5	0.3		412	348
31	9.1	4	119	110	1	51	63	39	0.0	0.0	4.6	1.0	505	393	362
32	5.9	100	59	1830	212	273	105	3220	0.7	3.4	1.1	12.6	10000	5817	6110
33	8.0	103	87	2220	238	302	42	3610	0.2	3.0	4.5	15.2	13000	6625	7300
34	9.8	0	66	168	20	330	224	40	0.2	0.9	9.9	0.9	1000	860	748
35	7.8	36	638	309	14	109	68	48	0.0	0.3	7.6	1.0	1250	1231	1198
36	6.6	72	101	211	24	52	686	36	23.7	0.7	3.8	0.4	1210	1211	864
37	6.8	61	74	120	36	76	471	21	11.7	0.3	2.1	0.9	850	874	637
38	6.7	63	73	109	35	67	477	20	12.3	0.3	1.8	26.3	850	885	624
39	7.2	32	58	148	22	60	419	28	7.5	0.5	2.6	10.3	815	788	576
44	5.5	105	62	1940	230	284	109	3310	0.7	3.6	0.7	12.8	9900		6040
46	9.8	1	77	184	22	331	269	48	0.2	1.1	11.1	1.0	845		831
47															
48															

Table 3.1, continued

LOC Abbreviation Key

LOC	BASIN NAME	LOC	BASIN NAME	LOC	BASIN NAME
---	-----	---	-----	---	-----
AMG	AMARGOSA DESERT	HVU	HAMLIN VALLEY	SCY	SILVER CITY
AVY	ANTELOPE VALLEY	IDA	IDAHO	SFB	SARCOBATUS FLAT
BDY	BRADY'S HOT SPRINGS	IMP	IMPERIAL VALLEY	SHM	SHEEP MOUNTAINS
BEO	BEOWAWE	JVY	JAKES VALLEY	SHR	SHEEP RANGE
BOI	BOISE, IDAHO	KBV	KOBEH VALLEY	SMA	STEAMBOAT-MOANA
BRD	BLACK ROCK DESERT	LKT	LAKE TAHOE	SMD	SMOKE DESERT
BSV	BIG SMOKY VALLEY	LSV	LITTLE SMOKY VALLEY	SMV	SMITH CREEK VALLEY
BTV	BUTTE VALLEY	LVC	LONG VALLEY CALDERA	SNR	SNAKE RANGE
BUF	BUFFALO VALLEY	LVS	LAS VEGAS SHEAR ZONE	SPM	SPRING MOUNTAINS
BVY	BUENA VISTA VALLEY	MDC	MODOC PLATEAU	SPR	SPRING VALLEY
CAL	CALIENTE	MIN	MINERAL COUNTY	SPV	SURPRISE VALLEY
CAR	CARLIN	MIS	MISCELLANEOUS	SSN	SOUTHERN SIERRA
CEV	CARSON-EAGLE VALLEY	MON	MONITOR VALLEY	STV	STEPTOE VALLEY
CLV	CLAYTON VALLEY	MRV	MUDDY RIVER VALLEY	SVY	SIERRA VALLEY
COL	COLADO	MSV	MASON VALLEY	SWI	SOUTHWEST IDAHO
COS	COSO	NEN	NORTHEAST NEVADA	TMD	TAHOE MEADOWS
CRF	CRATER FLAT	NTS	NEVADA TEST SITE	TSC	TUSCARORA
CSK	CARSON SINK	NWK	NEWARK VALLEY	UDV	UTAH DUGWAY VALLEY
CSR	COSO RANGE	OAV	OASIS VALLEY	UFS	UTAH FISH SPRING FLAT
CSV	COYOTE SPRING VALLEY	ORE	OREGON	UNC	UTAH NEWCASTLE
CVY	COAL VALLEY	PAH	PAHRANAGAT VALLEY	UPV	UTAH PINE VALLEY
DLV	DELAMAR VALLEY	PDV	PARADISE VALLEY	USC	UTAH SOUTH-CENTRAL
DPK	DESERT PEAK	PHM	PAHUTE MESA	USD	UTAH SEVIER DESERT
DXV	DIXIE VALLEY	PHS	PAINTED HILLS	USL	UTAH GREAT SALT LAKE
ECV	EDWARDS CREEK VALLEY	PHV	PAHRUMP VALLEY	USV	UTAH SNAKE VALLEY
ELK	ELKO AREA	PLK	PYRAMID LAKE	USW	UTAH SOUTH-WESTERN
FAL	FALLON AREA	PLV	PLEASANT VALLEY	UTL	UTAH LAKE
FLV	FISH LAKE VALLEY	PNE	PINE VALLEY	UTV	UTAH TOOLE VALLEY
FMC	FORTY MILE CANYON	PNV	PENOYER VALLEY	UWC	UTAH WEST-CENTRAL
FVY	FAIRVIEW VALLEY	PPV	PUMPERNICKEL VALLEY	UWV	UTAH WAH WAH VALLEY
GDV	GARDEN VALLEY	PRG	PAHRANAGAT RANGE	UWW	UTAH WHIRLWIND VALLEY
GSV	GRASS VALLEY	RBY	RUBY VALLEY	WKL	WALKER LAKE
GVY	GABBS VALLEY	RFT	RAFT RIVER IDAHO/UTAH	WKR	WALKER RIVER
HAW	HAWTHORNE	RNV	RALSTON VALLEY	WLS	WELLS
HCV	HOT CREEK VALLEY	RRV	RAILROAD VALLEY	WPR	WHITE PINE RANGE
HLV	HONEY LAKE VALLEY	RSE	REESE RIVER VALLEY	WRV	WHITE RIVER VALLEY
HUM	HUMBOLDT RIVER	RSV	ROSE VALLEY	WSV	WASHOE VALLEY
HVA	HIDDEN VALLEY	SCD	SMOKE CREEK DESERT	YCF	YUCCA FLAT
HVD	HOOVER DAM	SCV	STONE CABIN VALLEY	YCM	YUCCA MOUNTAIN

Elko Heat Company, the Warren Estates Geothermal Heating District, and Geothermal Food Processors. The remaining samples were taken from thermal springs throughout the study area.

In addition to fluid samples, three samples of carbonate scale (calcite) were obtained for radiocarbon age analysis. The scale had precipitated in the flash zone of power plant production wells at the Oxbow Dixie Valley Power Plant and Chevrons Desert Peak and Beowawe power plants. Although the scale is a contemporary mineral deposit, it is precipitated by geothermal fluids from a depth in excess of 3,000 m below the surface. The geothermal wells are cased isolating the fluids and preventing contamination from near surface modern groundwater. Radio-carbon age analysis should yield the age of the depositing fluid, not the time of deposition. The method of scale analysis is described in Appendix 1.

FINAL DATA BASE

The completed data base contains more than 1,200 entries; most include geochemical data and stable isotope data. The format adopted for the data base was taken from the Nevada GEOTHERM data base and includes the following information: sample name or site, reference, basin, source type, location (latitude and longitude), elevation, sample temperature ($^{\circ}\text{C}$), date, and chemical and isotopic data. During the course of the study, it was determined that the high temperature data was most useful. Figure 3.2 locates all data points above 80°C , from both this study and the literature search. The notable lack of high-temperature resources in some portions of

the study area presented some problems during interpretation, and is addressed in a later section.

STATISTICAL ANALYSIS OF DATA BASE

In order to determine the natural compositional variations among fluids within the study area, several statistical routines were developed and applied to the data base. Isotope ratios vary systematically, and statistical programs were used as a "first-cut" to estimate the magnitude of the variations. First, two programs, ISO-STAT and REGRESSN, were specifically developed for the study.

Figure 3.2: Distribution of data points greater than 80°C across the Great Basin. Note the lack of thermal resources in the south and northeast.

A third program, ABSTAT^R, is commercially available and was utilized for its multiple statistical operations and graphics capabilities.

Program ISO-STAT was designed to return basic statistics for isotope data. It was written on a FORTRAN^R 3.21 compiler and is designed to run on an IBM compatible computer. Statistical parameters include calculating the mean, standard deviation, and maximum/minimum values for both deuterium and oxygen-18. Analysis is performed on the regional data set, on seven one-degree latitude increment data subsets, and on thirteen individual drainage basins. Each data set is also separated into ten, 300 m elevation increment subsets. Program REGRESSN was designed to return correlation and statistics of oxygen-18 vs total dissolved solids (TDS). The program was also written using the FORTRAN 3.21 compiler and is IBM-PC compatible.

Program ABSTAT provides both statistical analysis and graphics for extremely large data sets (up to 4 megabytes), and effectively replaced the two fortran programs. The program was adapted to the existing data base, which is in a Lotus-123^R spreadsheet format. In addition to fundamental statistics, ABSTAT contains more exotic tests such as Kruskal-Wallis, rank correlations and cross correlations.

The initial correlations attempted during this investigation was the relationship between stable light-isotope ratios and total dissolved solids (TDS) of various geothermal fluids. The effort was based, in part, on the work of Clayton et al. (1966), from an

investigation of the origin of saline brines associated with hydrocarbon deposits in the Illinois, Michigan, Alberta (Canada) and Gulf of Mexico basins. They compared TDS with isotope content of the oil field brines in an attempt to determine the original water source. In most cases, extrapolating the brine isotope/TDS line lead directly to the plotting position of meteoric waters in the region. In a few cases, however, the extrapolated line did not intercept the position of the local meteoric water; the brines were more isotopically depleted than anticipated. In the absence of any appreciable elevated topography to provide a source of isotopically depleted waters, they concluded that recharge of the waters must have occurred during a cooler climate.

Using data from this investigation, a series of diagrams was generated comparing TDS to deuterium and oxygen-18. Examples of these diagrams are given as Figures 3.3 to 3.5. The preliminary analysis showed no clear trends. Carson-Eagle Valley in extreme western Nevada, and Paradise valley in northern Nevada (Figure 3.3) have a negative slope. Data from the Hoover Dam area in southern Nevada and Dixie Valley in west-central Nevada yielded a slightly positive slope (Figure 3.4). The Pumpernickel Valley area in northern Nevada and the Beowawe Area in eastern Nevada revealed essentially no slope (Figure 3.5). No clear overall trend could be established.

Figure 3.3: Deuterium vs. Total Dissolved Solids (TDS) for Carson-Eagle Valley, western Nevada, and Paradise Valley, northern Nevada. Slope is negative.

Figure 3.4: Deuterium vs. Total Dissolved Solids (TDS) for the Hoover Dam Area, southern Nevada, and Dixie Valley, north-central Nevada. Slope is positive.

Figure 3.5: Deuterium vs. Total Dissolved Solids for the Pumpernickel Valley, north-central Nevada, and the Beowawe area, eastern Nevada. Slope is roughly 0.

COMPUTER CONTOURS OF DATA SETS

The data sets were contoured using the Surfer contouring package from Golden Software, Inc. In a study of this type, where data points are at irregular intervals, adjusting the contouring parameters can produce drastically different results from the same data set. All plots presented here were contoured using the same parameters to avoid the introduction of any bias.

A grid setting of 15 established 225 equally spaced node points across a specified map area. An inverse distance weighting (IDW) routine calculated a value for each node point, using all points in the data set. IDW was used in favor of Kriging due to its greater overall speed and the similarity of results from test runs. "All" data points were used for each nodal calculation due to the irregularity of the data set distribution, and the above mentioned possibility of bias. No single search radius and method, or specified number of nearest points to use, adequately served all the data sets.

A liability of using the "ALL" data points setting is that contours will be drawn regardless of whether data exist in an area. Data points upon which the contours are drawn are provided with each plot to allow the reader to assess the validity of the contouring routine.

Stable Light-Isotopes

In order to assess any general trend in the data, contoured plots utilizing all available points, regardless of temperature or sample source, were generated. The plots of deuterium (Figure 3.6) and oxygen-18 (Figure 3.7) show the same general pattern. Since the fractionation of isotopes is temperature dependent, meaningful interpretation can come only when the data set is subdivided by temperature.

Non-Thermal Waters

Stable light-isotope content of precipitation is known to vary systematically on the basis of temperature, a function of latitude and elevation, and longitude, the "rain-out" factor (Dansgaard, 1964; Fritz and Fontes, 1980). See Appendix 2 for an explanation of "rain-out" and other characteristics of isotopes. Figure 3.8 demonstrates the close relationship between the worldwide meteoric water line (Craig, 1961) and the best fit line for non-thermal waters from the Great Basin on a deuterium vs. oxygen-18 plot of non-thermal springs from throughout the Great Basin. On this basis, it can be argued that high elevation cold springs recharged by precipitation should provide a reasonable estimation of the isotopic ratios in recent meteoric water. Since collection of precipitation for isotopic analysis is difficult, the above described data set will serve as the modern precipitation proxy. In order to avoid contamination due to regional interbasin flow (Mifflin, 1968), only springs discharging in the range, above the

Figure 3.6: Deuterium permil concentration for all points, regardless of sample temperature, throughout the Great Basin study area.

Figure 3.7: Oxygen-18 permil concentration for all points, regardless of sample temperature, throughout the Great Basin study area.

Figure 3.8: Deuterium vs. oxygen-18 for cold springs throughout the Great Basin. Best fit line is in good agreement with the worldwide Meteoric Water Line (Craig, 1961) defined as: $D = 8 \times ^{18}O + 10$.

Figure 3.9: Map of the western United States showing the predominant seasonal storm tracks of the Great Basin.

basin floor, were used for this determination. Cold wells were not used due to the same threat of contamination.

In the Great Basin, the anticipated distribution pattern of isotopes includes the effects of oceanic storm tracks and distance inland from the ocean. The two principal storm tracks that bring precipitation into the Great Basin (Figure 3.9) are northern storms that originate in the eastern Pacific Ocean, and southern storms that originate in the southern Pacific and Sea of Cortez.

In order to delineate the distribution of stable isotopes in meteoric waters, deuterium and oxygen-18 data from non-thermal springs (less than 20°C) were contoured on base maps of the Great Basin. Figure 3.10 shows the distribution of deuterium in these springs. The diagram illustrates the effects of latitude, and the winter and summer storm tracks. Specifically, deuterium content decreases (fluids become isotopically more depleted) from south to north reflecting the latitude effect, and from west to east due to the winter storm track "rain-out". Some of the contour configuration in the west may be attributed to the orographic effect of the Sierra Nevada. The most depleted meteoric waters (-128‰) are located in the extreme northeast corner of Nevada, while the most enriched fluids (-96‰) are located in southern Nevada.

Oxygen-18 shows a trend nearly identical to that of deuterium (Figure 3.11). The latitude effect produces a south-to-north gradient and the influence of the winter and summer track storms appears as a bulge in the contours. Values for these data range from -13‰ in the south to greater than -16‰ in the north.

Figure 3.10: Deuterium permil for non-thermal springs. Note trend of isotopic depletion from south to north and from west to east. Lack of data in southeast Oregon deflects contours west.

Figure 3.11: Oxygen-18 permil for non-thermal springs. Note similarity to the deuterium plot.

Thermal Waters

As this study pertains to geothermal fluids, it is important to assess the distribution of stable light-isotopes of high temperature fluids. Choosing a cut-off temperature for points to include in these plots proved to be a subjective process. The desire to exclude any fluids with a component of recent meteoric water had to be balanced with the necessity for a large enough population to lend credibility to the plots. After extensive analysis of the plots produced, 80°C was established as the minimum temperature for thermal waters.

Distribution of deuterium content of thermal springs (Figure 3.12) and wells (Figure 3.13) reveals a pattern similar to that for the non-thermal springs. Deuterium concentrations again become more depleted from north to south and from west to east. The only disruption of this pattern is an area of extreme depletion in the vicinity of Elko, Nevada. Oxygen-18 concentrations in thermal springs (Figure 3.14) and wells (Figure 3.15) show the same pattern.

Isotope variability, represented by non-thermal springs, is related to the prevailing meteoric conditions. The general pattern established does not appear to be affected by heating during deep circulation.

Figure 3.12: Deuterium permil for thermal springs. Note the pattern of isotopic depletion resembles that of the non-thermal springs, becoming more depleted from south to north and west to east.

Figure 3.13: Deuterium permil for thermal wells. Note the pattern of isotopic depletion resembles that of the thermal and non-thermal springs.

Figure 3.14: Oxygen-18 permit for thermal springs. Note similarity to the deuterium plot.

Figure 3.15: Oxygen-18 permit for thermal wells. Note similarity to the preceding plots.

Major and Minor Geochemical Constituents

In the process of determining the source of geothermal fluids in the Great Basin, it is both useful and necessary to examine not only the isotopic composition of the fluid, but the major and minor chemical constituents as well. Concentrations of the chemical constituents are derived from interaction with underlying rock units, usually at elevated temperatures, during deep circulation of the fluids. The following series of diagrams is intended to show the principal differences between dissolved chemical species and stable light isotopes.

Major Constituents

Sodium chloride (NaCl) rich geothermal fluids are among the most common throughout the world, and represent the highest temperature geothermal fluids found in the Great Basin. Two distinct geographical locations coincide with the highest concentrations of sodium chloride in geothermal fluids (Figure 3.16). The zone of high concentration in western Nevada is the Carson Sink, site of Pleistocene Lake Lahontan, while the other area is coincident with the site of Pleistocene Lake Bonneville, in Utah. The ultimate source of the sodium chloride fluid is believed to be from extensive rock-water interaction with the underlying marine sedimentary rocks, but incorporated playa-derived salts may play a role (Mariner et. al., 1983).

Sodium sulfate rich fluids are distributed similarly to the sodium chloride fluids (Figure 3.17). These fluids are believed to

Figure 3.16: Sodium + chloride (NaCl) concentration in ppm for all points greater than 80°C .

Figure 3.17: Sodium + sulfate (NaSO_4) concentration in ppm for all points greater than 80°C .

develop in association with sedimentary deposits that are rich in gypsum, but may also include volcanic areas that contain sulfide and sulfate ore minerals.

The distribution of carbonate-bicarbonate type fluids depends on well-depth, location, and temperature. Low temperature fluids ($<20^{\circ}\text{C}$), for example, exceed 1000 ppm bicarbonate and are coincident with the location of ancient Lake Lahontan, in northern Nevada. These high concentrations are the result of a combination of dissolution of carbonates along the flow path and evaporative concentration in the surface playas. The highest concentrations for geothermal fluids ($>80^{\circ}\text{C}$) occur in the Railroad Valley oil-field brines, in south-eastern Nevada (Figure 3.18). Oil wells in this area range from 1.0 to 2.5 km in depth and penetrate a thick wedge of Paleozoic limestones and dolostones, which account for the high carbonate-bicarbonate concentrations.

It is evident from the three fluid types discussed, there is very little correlation between the spatial distribution of stable light-isotopes and major chemical constituents of the geothermal fluids.

Minor Constituents

Silica is an important constituent in geothermal fluids as it can be used to estimate the temperatures of deep geothermal reservoirs (Fournier, 1981). Silica concentrations vary on the basis of temperature, pH, and availability of amorphous silica (Fournier and Rowe, 1966). Figure 3.19 shows the distribution of silica in

springs less than 20°C. The highest values appear to be associated with areas of recent volcanic activity. Lava flows can contain a significant amount of amorphous, highly soluble, silica in the form of obsidian and scoria. The concentration of silica in thermal springs and wells greater than 80°C (Figure 3.20) shows little correlation with the silica cold spring plot. The areas of high silica concentration correspond to the high temperature geothermal fields at Steamboat Springs, Wabuska and Beowawe, Nevada, and Long Valley, California.

Figure 3.18: Carbonate (CO_3) and bicarbonate (HCO_3) concentration in ppm for all points greater than 80°C.

Figure 3.19: Silica (SiO_2) concentration in ppm for all points less than 20°C.

Figure 3.20: Silica (SiO_2) concentration in ppm for all points greater than 80°C.

Boron is another minor constituent that is generally associated with geothermal activity. Concentrations of boron in samples less than 20°C (Figure 3.21) are limited to the western portion of the study area, in particular the Carson Sink. Distribution of boron in high temperature samples (Figure 3.22) is similar to that of silica, reflecting the location of geothermal fields.

As with the major constituents, there appears to be little correlation between minor chemical constituents and the distribution of oxygen and hydrogen isotope ratios.

RADIOACTIVE ISOTOPES

Analysis for two radioactive isotopes was performed for the purpose of age dating fluid samples. A complete description of the laboratory methods used and the theory of age dates obtained from radioactive isotopes can be found in Appendix 1 and Appendix 2, respectively.

Tritium

Sixteen fluid samples were collected for analysis of tritium, the radioactive isotope of hydrogen. With a half-life of 12.26 years, tritium has been used worldwide as a tracer and indicator of "modern" groundwater. The advent of above ground thermonuclear testing in the early 1950's flooded the atmosphere with bomb produced tritium.

Figure 3.21: Boron (B) concentration in ppm for all points less than 20°C.

Figure 3.22: Boron (B) concentration in ppm for all points greater than 80°C.

The usefulness of tritium analyses as an indicator of fluid age is limited to semi-quantitative age estimates. For example, the presence of tritium in amounts greater than a few tritium units (TU), indicates that some component of the fluid sampled has been in contact with the atmosphere since 1952. Conversely, an absence of tritium indicates no contact with the atmosphere since 1952. Tritium data are most useful used in conjunction with carbon-14 age dating, to indicate the presence of modern water that would invalidate the ^{14}C date.

Tritium data for all samples analyzed in the course of this study are listed in Table 3.2. All samples contain less than 5 TU tritium, the limit of detection for the procedure used. Typical values for surface water in northern Nevada range from 15 to 21 TU (Szecsody, 1982). The results indicate that no component of the fluid samples have had contact with the atmosphere since prior to 1952.

Table 3.2 - Tritium analyses, this study.

SAMPLE TU	LOCATION	SAMPLE TU	LOCATION
GF-1 <5	Brady's Hot Spring	GF-10 <5	Beowawe 1-13
GF-2 <5	Gerlach Hot Spring	GF-13 <5	Alakli Hot Spring
GF-3 <5	Patua Hot Spring	GF-14 <5	Bailey Hot Spring
GF-4 <5	Steamboat Hot Sp	GF-15 <5	Grapevine Sp, DV
GF-5 <5	Wabuska Tads #2	GF-18 <5	Rogers Sp, Lk Mead
GF-6 <5	Reno Warren Estates	GF-20 <5	Warm Springs
GF-7 <5	Soda Lake 33-14	GF-23 <5	Elko School Dist.
GF-8 <5	Desert Peak 67-21	GF-24 <5	Elko Heat Company

Carbon-14

An initial search of the literature revealed a significant data base of carbon-age data had been generated in the vicinity of the Nevada Test Site (NTS) and the proposed high-level radioactive waste repository at Yucca Mountain, Nye County. Hydrologic studies associated with the Site Characterization Plan (SCP) produced radiocarbon ages and stable light-isotope data for groundwaters in order to delineate recharge areas, flow rates, and the interaction between the water table and vadose zone. Additional data were obtained from geother-mal investigations in northern Nevada. Data were also obtained during the field sampling portion of this study.

Grove et al. (1969) determined a range of ages (0-23,000 years Before Present (BP) for groundwater from the NTS and speculated that the lakes that contributed this groundwater probably existed 12,000 to 20,000 years BP. In the Amargosa Desert, Claassen (1985) determined that groundwater was recharged primarily by overland flow of snow melt in stream channels during the late Pleistocene. White and Chuma (1987) used stable isotopes and radiocarbon ages to delineate possible recharge areas and flow rates for groundwater systems in the Oasis Valley-Fort Mile Canyon basins in southern Nevada. Various corrections were applied and the calculated ages ranged from 3,000 to 12,000 years BP. Lyles and Hess (1988) used a combination of ion, stable light-isotope and radiocarbon ages to delineate the extent of the Las Vegas Shear Zone in southern Nevada. Geochemical traverses both parallel to and across the zone identified a range of fluid ages from 2,500 to 12,500 years BP.

Glancy (1986) used stable isotopes and radiocarbon data to help in the delineation of groundwater systems in the Fallon area, northern Nevada. Adjusted ages range from modern to 8,800 years. Recharge was attributed to inflow from the Carson River from the Lahontan Reservoir. Additional data were derived from geothermal investigations in Dixie Valley (Jacobsen et al., 1983) and the Steamboat Moana Area (Flynn and Ghusn, 1984).

Fluid samples were collected during the course of this investigation at the sites identified in Figure 3.23. In addition, carbonate scale was obtained from geothermal production wells in

Figure 3.23: Location of high temperature carbon age dates used in this study.

Dixie Valley (Oxbow Geothermal), Beowawe and Desert Peak (Chevron Resources). The details of collection, preparation and analysis are given in Appendix 1. Radiocarbon data from this study are listed in Table 3.3.

Table 3.3 - Carbon-age data from this study. IC indicates there was insufficient carbon present to complete the analysis.

GF #	SAMPLE NAME	LONG	LAT	^{14}C pmc	$^{13}/^{12}\text{O}/\text{OO}$	UNCORR AGE
32	Soda Lake 33-14	118.86	39.56	IC	IC	
33	Desert Peak Inj.	118.95	39.75	IC	IC	
34	Beowawe Batz Inj.	116.62	40.55	6.1	-0.7	23100
36	Warm Spring	116.38	38.19	<0.6	-0.6	>42300
37	Elko School Dist.	115.76	40.84	<1.6	-0.8	>34200
40	Beowawe 1-13 Scale	116.62	40.55	8.3	-4.0	20200
41	Oxbow 45-5 Scale	117.94	39.95	2.1	-5.1	30900
42	Oxbow 74-7 Scale	117.95	39.94	2.4	-5.8	29800
43	Oxbow 76-7 Scale 4	117.95	39.94	3.7	-6.2	26500
46	Beowawe 2-13	116.62	40.55	8.7	-4.4	20500
48	Wabuska Tads #2	119.15	39.16	IC	IC	
49	Desert Peak Scale	118.95	39.75	<0.5	-2.2	>43800

CHAPTER 4. PALEOCLIMATE PROXY DATA BASE

ARCHAEOLOGICAL DATA SET

American Indian artifacts are widespread throughout the Great Basin, and documented carbon-14 age determinations indicated that several tribes inhabited portions of Nevada from 10,000 years BP to the present. Artifacts of particular interest are plant material and reed baskets found in food and natural resource caches.

Archaeological and other paleoclimate proxy data that were collected during the course of this investigation are shown in Table 4.1. The data were divided into four major categories, each representing a different proxy type: pollen (55 entries); charcoal, peat and blackmat sediments (33 entries); Neotoma species middens (98 entries); and gastropod and bone material (97 entries). The compilation illustrated initially that a sufficiently varied proxy data base was available to provide the investigation with stable isotope and carbon-14 age dates throughout the Great Basin.

Upon closer review of the archaeological data, two related problems emerged. Although the data may be successfully used to document the inhabited site, from an archaeological perspective, there was no guarantee that the plant material was indigenous to the site. In fact, on the basis of the nomadic activity of the tribes that inhabited the area, in all likelihood it was not. In addition to questionable location data, there was no control on elevation of the sample. Since deuterium is known to vary with both location and elevation, and in the absence of specific

Table 4.1 - Archaeological sites in the Great Basin.

P - Pollen N - Neotoma, midden studies G - Gastropods, dung, bone, C - Charcoal, peat, blackmat seds

SITE	REFERENCE	AGE	+/-	LAT	LONG	SITE	REFERENCE	AGE	+/-	LAT	LONG
G 1	9	9053		43.38	119.79	G49	31	10085	400	41.92	115.40
G 2	9	11000		42.65	120.13	G50	9	9500		41.79	114.34
G 3	17	25280	750	41.39	118.13	G51	9	5400		41.81	113.87
G 4	17	875	711	40.78	119.49	G52	51	12000		41.91	113.41
G 5	17	> 40000		40.82	118.41	G53	26	540000		41.75	112.04
G 6	8	28790	1010	40.64	118.00	G54	40	7080	160	41.74	111.79
G 7	17	13430	195	40.55	119.44	G55	7	15800	400	41.62	112.48
G 8	17	17170	270	40.38	119.85	G56	36	15800	400	41.37	112.53
G 9	36	10500	200	40.41	119.22	G57	36	21200	1000	41.32	112.50
G10	36	17300	400	40.32	119.18	G58	7	3730	350	41.16	112.21
G11	36	17900	800	40.29	119.86	G59	36	9800	300	41.03	111.92
G12	17	19620	350	40.27	119.81	G60	43	25300	1000	40.81	112.37
G13	43	17500	600	40.20	119.79	G61	40	14410	110	40.87	111.82
G14	43	14500	400	40.16	119.74	G62	9	14150	800	40.85	111.78
G15	43	19750	650	40.22	119.44	G63	9	11690	190	40.82	111.83
G16	43	10700	240	40.20	119.42	G64	40	19760	20	40.77	111.90
G17	9, 12	11200	250	40.19	119.40	G65	40	14150	800	40.74	111.93
G18	43	12700	300	40.02	119.54	G66	7	10300	310	40.70	112.12
G19	43	28900	1400	39.95	119.54	G67	43	23150	100	40.63	112.19
G20	43	11250	350	39.92	119.64	G68	40	11690	190	40.65	111.98
G21	22	19800	300	39.93	119.51	G69	40	9930	210	40.68	111.94
G22	17	40000		39.92	119.33	G70	24	740000		40.66	111.54
G23	42	> 40000		39.85	119.40	G71	7	11300	250	40.53	113.67
G24	43	> 34000		39.77	119.34	G72	36	11400	300	40.03	113.07
G25	17	21480	370	39.85	118.91	G73	36	18000	400	39.98	112.95
G26	36	28500	1700	39.86	118.54	G74	43	23300	800	39.93	112.86
G27	30	24500		39.86	118.33	G75	43	33200	4000	40.04	111.84
G28	30	33650	1720	39.76	118.50	G76	36	19300	400	39.88	111.88
G29	17	22275	175	39.74	118.45	G77	36	17000	400	39.56	112.34
G30	8	8800	90	39.64	119.31	G78	36	11600	200	39.50	112.41
G31	17	18900	340	39.60	119.23	G79	36	17700	420	39.45	112.45
G32	33	17600	650	39.26	120.19	G80	36	15000	500	39.45	112.37
G33	8	8680	200	39.60	118.73	G81	36	18000	4000	39.32	112.43
G34	30	11500	300	39.51	118.96	G82	47	23000		39.46	114.00
G35	42	3500	195	39.48	118.80	G83	10	6120		38.85	114.26
G36	36	13600	400	39.51	118.61	G84	7	10360	225	39.01	113.03
G37	36	11800	200	39.53	118.51	G85	9	17000		37.03	116.36
G38	36	11600	200	39.38	118.88	G86	28	14300	800	36.28	115.24
G39	30, 72	> 40000		39.39	118.75	G87	45	> 40000		36.22	115.20
G40	53	21100		39.36	118.73	G88	9	13100	200	36.22	115.14
G41	36	> 35000		39.45	118.58	G89	9, 23, 46	11690	250	36.17	114.88
G42	36	> 40000		39.41	118.58	G90	27	11010	400	36.11	114.54
G43	8, 14, 38	11880	160	39.19	119.24	G91	23	40000		35.97	113.97
G44	2	9980	800	38.87	120.05	G92	24	33600	1000	35.88	113.73
G45	22	3340	180	38.68	118.74	G93	27	11870		36.08	113.61
G46	8	11880	160	38.61	118.73	G94	23	11290	170	36.11	113.29
G47	41	45000		38.58	118.75	G95	34	11860	95	35.22	116.12
G48	9, 32	5300		38.71	116.89	G96	6	15350	240	35.24	116.04

Table 4.1 - continued. Archaeological sites in the Great Basin.

SITE	REFERENCE	AGE	+/-	LAT	LONG	SITE	REFERENCE	AGE	+/-	LAT	LONG
G97	35	10270	160	35.21	116.00	N50	4,1	19620	600	35.08	114.53
N 1	33	12020	470	40.36	119.32	N51	5	5800	250	34.78	116.64
N 2	36	12070	400	40.31	119.27	N52	4,48	11850	550	34.65	116.74
N 3	44	6400		41.48	113.20	N53	5	8300	700	34.41	116.90
N 4	32	13620	700	39.77	113.91	N54	19	10420	350	34.67	115.62
N 5	32,47	27280	970	39.47	114.23	N55	19	10460	330	34.56	115.59
N 6	32	11600	245	39.45	114.15	N56	37	17000		34.55	114.37
N 7	32	17350	435	39.43	114.09	N57	4,13	19500	380	34.12	114.94
N 8	32	12230	180	38.95	113.89	N58	4,13	13900	200	34.09	114.93
N 9	32	13480	250	38.92	113.89	P 1	2	> 50000		39.19	120.16
N10	13,39	9450	90	37.57	116.25	P 2	2	110000		38.89	120.14
N11	13	8420	100	37.56	116.14	P 3	2	> 50000		38.92	120.00
N12	32	20000	400	37.58	114.57	P 4	2	> 10000		38.93	119.94
N13	32	14980	250	37.52	114.57	P 5	2	15565	820	38.07	120.06
N14	1	27280	950	42.60	117.16	P 6	1	11350	650	37.82	118.67
N15	38	9320	300	36.88	116.20	P 7	7	14260	100	40.46	112.37
N16	38	16800	300	36.89	115.93	P 8	1	10200	150	39.48	114.37
N17	38	10100	160	36.86	115.92	P 9	1	10525	245	39.45	114.37
N18	9	12700	200	36.71	115.94	P10	1	30365	550	39.42	114.37
N19	38	9000	250	36.68	115.93	P11	1	11025	230	39.40	114.39
N20	1,13	11600	160	36.39	116.67	P12	1	37000		39.37	114.40
N21	19	10260	520	36.40	116.27	P13	1	21000	450	39.37	114.25
N22	19	14810	400	36.53	116.09	P14	1	35600	3000	39.40	114.15
N23	19	11680	650	36.56	116.07	P15	1	11785	270	38.87	114.61
N24		18680	280	36.55	115.37	P16	1	36300	300	38.96	114.52
N25	4	18800	130	36.57	115.15	P17	1	11800	330	39.06	114.40
N26	4	17420	250	36.54	115.18	P18	1		155	39.00	114.32
N27	4	19400	300	36.49	115.19	P19	1	> 31400		39.04	114.31
N28	4	21350	440	36.49	115.14	P20	1	11895	270	39.03	114.24
N29	4	11550	150	36.47	115.18	P21	1	10370	150	39.07	114.11
N30	32	20380	340	36.49	115.27	P22	1	11880	170	39.16	113.72
N31	32	21350	440	36.46	115.27	P23	1	10340	150	36.12	113.72
N32	4	21700	500	36.44	115.32	P24	1	12160	1800	39.06	113.64
N33	4	20380	340	36.41	115.34	P25	1	12900	650	38.67	113.70
N34	16	31400	1500	36.28	115.15	P26	1	12470	290	38.64	113.66
N35	16	21700	500	36.26	115.19	P27	1	12830	285	38.56	113.68
N36	16	22000		36.20	115.19	P28	1	10485	235	38.55	113.57
N37	5,28	31300	250	36.20	115.14	P29	1	12570	180	37.59	114.31
N38	4	13800	400	35.53	117.89	P30	1	20100	850	37.57	114.34
N39	32	14900	180	35.90	115.62	P31	1	2170	120	37.48	114.43
N40	1,4	15040	650	36.05	115.37	P32	1	9550	650	36.83	117.36
N41	4	15800	680	36.03	115.35	P33	1	28900	1200	36.72	115.99
N42	15	> 40000		36.03	113.90	P34	1	16400	250	36.72	115.53
N43	1,4,39	12650	380	36.00	113.83	P35	2	8000		36.45	117.81
N44	4	14810	220	35.97	114.74	P36	1,6	13150	500	36.44	116.21
N45	4	16580	460	35.70	113.35	P37	1	> 40000		36.56	115.93
N46	16,32	28720	800	35.57	115.65	P38	1	17750	200	36.58	114.43
N47	16,29	23600	1500	35.50	115.65	P39	1	15400	205	36.16	114.68
N48	34	15350	240	35.35	116.07	P40	1,39	6330	270	35.96	113.87
N49	54	13380	300	35.26	114.76	P41	1,39	16580	460	36.16	113.32

Table 4.1 - continued. Archaeological sites in the Great Basin.

SITE	REFERENCE	AGE	+/-	LAT	LONG	SITE	REFERENCE	AGE	+/-	LAT	LONG
P42	2	5000		35.94	117.90	C11	14	19990	380	39.89	118.29
P43	2	13800	400	36.63	117.98	C12	14,12	11199	570	40.02	118.53
P44	2	32000		35.74	117.48	C13	12	8130		39.13	120.25
P45	1	11525	150	35.74	114.50	C14	14	11275	175	39.09	118.61
P46	1	12625	150	35.14	114.66	C15	14	21480	175	38.78	118.89
P47	1,13	9140	140	34.47	116.81	C16	14	12240	160	38.67	118.89
P48	1	12100	400	34.46	116.75	C17	14	25280	750	38.75	118.63
P49	1	16900	190	34.74	114.54	C18	7	12090	300	42.33	112.00
P50	1	19500	380	34.45	114.73	C19	7	26700	900	41.49	113.25
P51	1	13900	200	34.41	114.71	C20	3	10695	310	41.42	113.06
P52	1	12960	210	34.23	114.29	C21	3	6000	225	41.58	112.73
P53	1	12330	350	34.18	114.26	C22	7	20800	400	41.63	112.51
P54	1 >	30000		34.23	113.52	C23	7	11453	600	40.68	114.00
P55	1	10250	200	34.15	113.53	C24	7	26300	1100	40.93	112.38
C 1	9	11200		43.52	119.75	C25	3	10285	265	40.59	112.21
C 2	9	13200		43.33	119.78	C26	7	17580	270	40.57	111.78
C 3	9	9500		42.63	117.40	C27	3	26080	1200	40.56	111.66
C 4	8	22000		41.23	118.68	C28	3	12300	330	40.61	111.55
C 5	9	6500		41.00	119.42	C29	3	9430	260	40.53	111.58
C 6	49	2465	100	40.53	120.80	C30	7	23600	1000	39.34	112.60
C 7	18	33650	1720	40.79	119.19	C31	20	296000	116k	36.95	118.24
C 8	14	21370	420	40.27	119.64	C32	2	2000		37.05	117.84
C 9	8	33650	1720	39.89	118.40	C33	11	1300		36.19	115.93
C10	43	9720	100	39.76	118.38						

Archaeological References

#	YEAR	REFERENCE	#	YEAR	REFERENCE	#	YEAR	REFERENCE
1	1983	Wells	18	1983	Davis	36	1965	Broecker and Kaufman
2	1985	Bryant and Halloway	19	1983	Spaulding	37	1979	Wells
3	1980	Curry	20	1982	Dalrymple et al	38	1964	Wells
4	1979	Van Devender and Spaulding	21	1986	Feduccia and Oviatt	39	1976	Wells
5	1977	Van Devender	22	1988	Lao and Benson	40	1980	Nelson and Madsen
6	1982	Weide	23	1974	Long and Martin	41	1988	Benson
7	1982	Curry and James	24	1981	Mead and Phillips	42	1965	Morrison and Frye
8	1982	Davis	25	1972	Madsen	43	1958	Broecker and Orr
9	1982	Grayson	26	1987	Oviatt and McCoy	44	1970	Aikens
10	1982	Thomas	27	1984	Phillips	45	1967	Haynes
11	1982	Lyneis	28	1986	Quade	46	1970	Heizer and Berger
12	1982	Elston	29	1977	Spaulding and Van Devender	47	1985	Heaton
13	1967	Wells	30	1984	Spencer	48	1976	King
14	1981	Benson	31	1963	Shutler	49	1979	Davis and Pippin
15	1974	Phillips and Van Devender	32	1982	Thompson and Mead	50	1972	Born
16	1977	Spaulding	33	1986	Thompson et al	51	1977	Merringer
17	1978	Benson	34	1987	Wells et al	52	1985	Davis
			35	1978	Warren and Ore	53	1985	Grayson
						54	1970	Leskinen

controls, it was determined that any stable isotope data derived from this source could be misleading.

PACKRAT MIDDEN SAMPLES

In this investigation, 25 samples of juniper twigs collected from ancient and modern packrat middens at five different sites in northern and central Nevada were obtained from Dr. Peter Wigand of the Desert Research Institute, Reno, Nevada. In addition to the midden material, Dr. Wigand provided radiocarbon ages for all samples. Packrat midden stable light-isotope and radiocarbon age data used in this study are listed in Table 4.2, midden sample locations from this investigation are shown in Figure 4.1.

Midden preparation includes several cleaning, soaking, and separation processes that isolate various kinds of plant macrofossils. The botanical studies consisted of radiocarbon dating of selected material and descriptions of the entrained plant species. Deuterium analyses are performed on cellulose that is separated from the juniper twigs. A complete discussion of the process is included in Siegel (1983).

In practice, twigs from Utah juniper are preferred because of its widespread temporal and spatial distribution throughout the western United States. This abundance is reflected in the high volume percentage of this material in the middens collected. The plant species selected should be in sufficient supply within all middens selected for deuterium analysis.

Table 4.2 - Current study, packrat middens analyses

SAMPLE	LOC	LONG	LAT	DEUT o/oo	ISO MAT	CARB AGE	+/-	DATE MAT
SC110388PEW1	SCY	119.64	39.26	-60	J	5650	150	N
SC111389MJ1	SCY	119.64	39.26	-55	J	MJ		
HV200589PEW1	HVY	119.68	39.49	-57	J	9800	340	N
HV111289MJ1	HVY	119.68	39.49	-48	J	MJ		
SM110389CLN2(1)	SMD	119.82	40.32	-64	J	11790	130	N
SM110389CLN3(1)	SMD	119.83	40.35	-73	J	10810	150	N
PH180589PEW2(1)	PHS	119.66	39.88	-64	J	21460	420	N
PH140889PEW2(1,5)	PHS	119.64	39.88	-71	J	24580	470	N
PH180589PEW3(2,1)	PHS	119.64	39.88	-71	J	16280	160	N
PH140889PEW2(1,3)	PHS	119.64	39.88	-79	J	11910	110	N
PH140889PEW2(1,4)	PHS	119.64	39.88	-49	J	11530	160	N
PH140889PEW2(1,2)	PHS	119.64	39.88	-59	J	8560	120	N
PH211089PEW1(1,2)	PHS	119.64	39.88	-65	J	29630	590	J
PH211089PEW1(1,3)	PHS	119.64	39.88	-55	J	28560	390	J
PH211089PEW1(1,4)	PHS	119.64	39.88	-53	J	11730	100	J
PH211089PEW1(1,5)	PHS	119.64	39.88	-77	J	23110	410	J
PH211089PEW1(1,6)	PHS	119.64	39.88	-40	J	29410	770	J
PH211089PEW1(1,7)t	PHS	119.64	39.88	-89	J	12440	130	J
PH211089PEW1(1,7)m	PHS	119.64	39.88	-77	J	12340	190	J
PH211089-MJ	PHS	119.64	39.88	-71	J	MJ		
PR100588PEW1(4,1)	PRG	115.40	37.41	-38	J	9540	120	
PR100588PEW1(4,2)	PRG	115.40	37.41	-34	J	5820	90	
PR100588PEW1(5,1)	PRG	115.40	37.41	-52	J	11710	150	
PR100588PEW1(5,3)	PRG	115.40	37.41	-61	J	22730	1080	
PR101189MJ1	PRG			X	J	MJ		

LEGEND

HVA	- Hidden Valley, Sparks Nevada	S	- Symphoricarpos
PRG	- Paharanagat Range, Southern Nevada	P	- Pinus longaeva
SMD	- Smoke Creek Desert, Northwest Nevada	PM	- Pinus Monophylla
PHS	- Painted Hills, North of Reno	MJ	- Modern Juniper
SCY	- Silver City, Southeast of Reno	J	- Juniperus
		N	- Fecal pellets

ISO - Material Analyzed for isotope content
MAT

CARB - Sample age in radiocarbon years
AGE

DATE - Material analyzed for carbon age
MAT

Figure 4.1: Packrat midden sample locations. Number of samples taken at each location shown in parentheses. SR=Snake Range, SC=Silver City, PR=Pahranagat Range, SM=Smoke Creek Desert, HV=Hidden Valley, PH=Painted Hills.

CHAPTER 5. GEOTHERMAL RECHARGE MODELS

HISTORICAL MODELS

The origin of geothermal fluids has been the subject of interest and speculation for more than a century. Dana (1883) suggested that the action of geysers in Yellowstone Park "...is owing to the access of subterranean waters to hot rocks, producing steam, which seeks exit by conduits upward". Ironically, he identified all three components of a geothermal system simply by observation, at least twenty years before the world's first commercial power plant went on line in Larderello, Italy. Investigators today rely principally on geochemical survey techniques, but the components remain unchanged.

The advent of geothermal fluid recharge models is associated with the rapid development of geothermal energy resources in the western United States that began in the mid 1970's. Geothermal fluids were known to be chemically and physically different from non-thermal surface and groundwater, but their source and flow path would remain largely speculative until a chemical tracer could be identified. In 1963, Craig identified the tracer, naturally occurring stable light-isotopes of hydrogen and oxygen.

On the basis of isotopic ratios alone, Craig demonstrated that most geothermal fluids throughout the world were of a meteoric origin. However, comparison of isotopes of high temperature geothermal fluids with local non-geothermal fluids revealed two important distinctions: the geothermal fluids were, 1) isotopically depleted

(lighter) with respect to deuterium, and 2) isotopically enriched (heavier) with respect to oxygen-18. This section discusses the application of isotopes to explain geothermal fluid genesis.

Oxygen participates in chemical reactions with rocks much more readily than hydrogen, and is not a good candidate for tracing geothermal fluid origin. Hydrogen and its isotopes have been shown to be significantly less reactive and are overwhelmingly preferred for studies of this type. It should be noted that the models presented in the following discussion are based principally on deuterium concentrations.

On the basis of deuterium content, two schools of thought have emerged that explain the origin of thermal fluids: recent recharge and paleo-recharge. The data set for both is the same, only the assumptions and interpretations differ.

Fournier and Thompson (1980) concluded that the most probable recharge areas for isotopically depleted geothermal fluids at Coso Hot Springs in eastern California, were the slopes above 1,980 m elevation in the distant Sierra Nevada. They observed that the local precipitation was isotopically heavier than the Coso thermal fluids. Faults and fractured rock were identified as the principal conduit for geothermal fluids.

North of Coso at Steamboat Hot Springs, Nevada, Nehring (1980) cited isotopic concentrations in suggesting that recent meteoric fluids were recharging the hydrothermal system at an estimated elevation of 1,920 m above sea level. The most likely source was

the Carson Range to the west, but a contribution from the Virginia Range to the east, could not be ruled out. Bell et al. (1980) examined thermal fluids from wells and springs in Dixie Valley, Nevada and established two separate recharge areas, one above and the other below 2100 m elevation. The recharge zones produced two distinct thermal fluids with separate flow paths, which then mixed in various proportions to produce the sampled fluids. Jacobson et al. (1983) reached a similar conclusion, and suggested an age for the Dixie Valley thermal fluids of 2,000 to 5,000 years BP.

In the central Great Basin, Bowman and Cole (1983) determined that the Independence Mountains were the source of recharge for the hydrothermal fluids associated with the Tuscarora thermal area. A closer highland area was ruled out as the source because it did not contain sufficiently depleted non-thermal fluids. Geophysical evidence also suggested fluid flow in the area trended from the southeast to northwest.

In the eastern Great Basin, Rohrs and Bowman (1980) developed a model for geothermal fluid recharge for the Roosevelt Hot Spring hydrothermal system, Utah. Citing stable light-isotopes of recent precipitation, they argued that the Mineral Mountains above 1980 m elevation was the likely source of fluid recharge. Cole (1983) investigated thermal springs throughout Utah and developed a model for geothermal recharge based on correlations with non-thermal, high elevation precipitation. Non-thermal fluids from the Wasatch Front provided the best correlation and allowed for an age estimation of 13 to 550 years BP.

In their assessment of the hydrothermal system of Leach Hot Springs, in Pershing County, Nevada, Welch et al. (1981) found no nearby "suitably elevated" recharge area for the local isotopically depleted geothermal fluids. They suggested that either the lateral flow path was at least 160 km (distance to the nearest isotopically depleted source), or the waters were "paleowaters" that had fallen during a colder, wetter climatic period. They also compared two conceptual fluid recharge models on the basis of geochemical and heat flow data. The fault plane model, in which fluid circulation is confined to discrete portions of a single fault, was judged non-viable on the basis of the relatively short calculated fluid residence time. A second concept, the lateral flow model, was based on recharge along one fault zone, lateral flow beneath the valley, and discharge up a fault zone on the opposite side of the valley. This model was considered more tenable because the estimated residence time was consistent with the isotope data.

Welch and Preissler (1986) evaluated ion and stable isotope aqueous geochemistry at Brady's Hot Springs, Churchill County, Nevada, and concluded that local meteoric water, concentrated by evaporation, was the most likely source of the high TDS geothermal fluids.

Mariner et al. (1983) reviewed the chemical and isotopic composition of approximately 150 geothermal springs in the northern Basin and Range province. They found that the difference in deuterium content between modern, meteoric waters and the thermal fluids was "almost certainly an indication that the thermal waters

probably recharged during times of colder climate, probably the Pleistocene."

SUMMARY OF MODELS

Surveys of geothermal systems have demonstrated that thermal fluids are consistently more depleted in stable light-isotopes than non-thermal fluids discharging at comparable elevations (Trexler et al., 1979, 1980, 1981, 1982; Flynn et al., 1984). Interpretations of these isotopic differences include contemporary recharge from adjacent mountains (Nehring, 1980; Rohrs and Bowman, 1980; Fournier and Thompson, 1980; Cole, 1983) and recharge during pluvial/glacial times (Welch et al., 1981; Mariner et al., 1983). In a study of the Beowawe system, Day (1987) pointed out that geothermal systems in northern Nevada are systematically distributed in a pattern similar to the isotopic concentrations of modern precipitation in the Great Basin.

The key to the resolution of these differing interpretations was a detailed analysis of prevailing paleoclimatic conditions in the Great Basin during the Pleistocene. Of particular importance is the extent and magnitude of cooler, wetter conditions that would further an isotopically depleted precipitation regime.

CHAPTER 6. PALEOCLIMATOLOGY

GLOBAL STUDIES

The models discussed in the previous section that assume that geothermal fluid recharge is a recent event, fail to account for the significant change in the isotopic composition of precipitation over the last 40,000 years. Ironically, it has been known for over twenty years that the isotopic composition of precipitation varies as a function of climate and that climate varies as a function of time. The magnitude and location of those variations are the subject of the next section.

Perhaps the most compelling evidence for temporal variations in isotopic composition was presented by Dansgaard et al. (1969). Using ice from a core-hole at Camp Century that penetrated the entire thickness of the Greenland Ice Sheet (1390 m), the variation in the isotopic content of precipitation for the last 15,000 years was determined. In addition to the anticipated annual variations, they discovered that modern precipitation is significantly more enriched in stable light-isotopes than older precipitation. The transition from ancient, depleted fluids to modern, enriched fluids took place over a 2,000 year period of time between 11,000 and 9,000 years BP. This time period corresponds to the last major retreat of the continental glaciers. The authors found that the timing of the Oxygen-18 vs. ice stratigraphic age curve was in general agreement with known climate change timing in other parts of the world. Subsequently, Johnson et al. (1970) used spectral analysis to identify systematic, long-term climatic oscillations

with periods of 400 and 24,000 years from the ice core data. Following the work at Camp Century, additional core holes were completed to various depths throughout Greenland. The deepest of these is Dye 3, drilled to bedrock in south Greenland to a depth of 2,037 m. A comparison of oxygen-18 values between Camp Century and Dye 3 (Dansgaard et al, 1985) shows a remarkable correlation in the timing of isotopic enrichment and depletion events over the last 90,000 years.

Martinson et al. (1987) developed a "high-resolution" chronostratigraphy based on the oxygen isotope ratio of tests from benthic foraminifera (forams). The procedure utilized the "standard" deep-sea sediment stratigraphy of Pisias et al. (1984), five climatically responsive data sets, and four separate orbital tuning methods to establish the chronology. According to the theory of orbital tuning, originally developed by Milankovitch (1941), climatic changes, including glaciation, result from solar radiation fluctuations determined by variations of the Earth's orbital elements and axial inclination. Figure 6.1 demonstrates the close correlation of the curves drawn from the Dansgaard and Martinson studies. The curves are presented on a normalized scale to allow easy comparison of the different data types, ice and foraminifera, respectively.

GREAT BASIN STUDIES

The Dansgaard and Martinson studies clearly demonstrate the depletion in stable light-isotopes of meteoric waters that occurred

during the last glacial period. The location of the two studies at opposite poles of the globe suggest the event was worldwide, but was the depletion recorded in the Great Basin?

Devils Hole, Amargosa Desert

Devils Hole is a 1 to 5 m wide, open fault zone in the modern spring lineament at Ash Meadows, Amargosa Desert, Nevada. The Hole extends more than 100 m below the water table, which is about 15 m below ground surface. Winograd et al. (1988) analyzed a calcite (CaCO_3) vein found approximately 21 m below the water table. The absence of dissolved Thorium (<0.01 ppb) in the groundwater allowed dating of the layers in the calcite vein by the $^{230}\text{Th}/^{234}\text{U}$ method. Oxygen isotope analysis of the dated vein layers allowed plotting of a paleoclimate curve similar to that of Martinson et al. (1987). A comparison of these curves is shown in Figure 6.2, and the agreement is good. The phase difference in the Winograd curve with respect to Martinson's ranges from about 17,000 years at point A (Figure 6.2), and increasing linearly to 25,000 years at point H. Winograd et al. (1988) attributed the difference to inaccurate age assignments for the Martinson study.

Unfortunately, deposition of the calcite vein ceased at around 51,000 years BP, the period of interest for this study. The reason for this curtailment is unclear, as the groundwater remains super-saturated with respect to calcite and precipitation should occur. The record obtained does confirm, however, that Great Basin groundwater records worldwide isotopic trends.

Figure 6.1: Planktonic ^{18}O vs. stratigraphic age from sub-polar Indian Ocean deep sea cores (Martinson et al., 1987), compared to ice ^{18}O vs. ice calculated age from Greenland ice cores (Dansgaard et al. 1969). Timing of changes in isotopic content is in good agreement.

Figure 6.2: Comparison of Martinson et al. (1987) data with ^{18}O of groundwater recorded in calcite at Devils Hole, Amargosa Desert, Great Basin (Winograd et al., 1988). Letters on the curves represent major peaks and troughs thought by Winograd to be correlative.

Glaciology

Several snowfields and small glaciers in the Sierra Nevada are thought to have persisted since the last major glacial event in the Great Basin. Temporally continuous glaciers and ice fields at various locations have provided paleoclimatological data for the past 100,000 years (Dansgaard et al., 1969). However, North American glacial investigations deal principally with the differentiation and classification of deposits of first order glaciation (Porter et al., 1983). No record of isotopic analysis of Great Basin ice cores was discovered.

Blackwelder (1934) was one of the first to describe the distribution and intensity of Great Basin glaciers and determined that they were influenced by three factors: latitude, humidity, and altitude. The Ruby Mountains (top elevations of 3,200 to 3,400 m) in the northeastern portion of the state, for example, supported vigorous glaciers with several stages of advance and retreat. On the other hand, the Spring Mountains in southern Nevada attain an elevation of nearly 3,600 m, but apparently did not receive sufficient precipitation to support glaciers. Likewise, the Sierra Nevada was much more heavily glaciated than the White Mountains, immediately to the east, due to the fact that the Sierra receives more precipitation. The White Mountains attain an elevation of 3,900 m, but are in the rain shadow of the Sierra.

Curry (1969) developed a model of glacial activity in the Great Basin that is based on radiocarbon ages, lichenometry, dendrochronology and other data. He identified four time periods of

increased snowfall that resulted in four periods of multiple glacial advance during the last 10,000 years: 6,000 to 7,000 years BP; 2,000 to 2,600 years BP; 1,000 years BP; and 650 years to 60 years BP. Assuming a 40 to 50 percent increase in seasonal snowfall and associated cooler and cloudier summers, he concluded that all of the climatic variations thought to have occurred in the Sierra Nevada during the last 10,000 years could have been the result of climatic variations with the magnitude of those observed during the past 100 years.

LaMarche (1965) mapped the distribution of glacial features in the White Mountains of California and Nevada and concluded that only the eastward-draining canyons were glaciated. To the east, Wayne (1982) restudied alpine glaciation in the Ruby Mountains and East Humboldt range and concluded that only one deposit, Angel Lake, could be correlated to the Late Wisconsin age. Neither study provides more than qualitative information on paleoclimates in the Great Basin.

Dohrenwend (1984) reviewed the glacial geomorphological data for the Great Basin and determined, on the basis of "nivation" landforms, that mean annual temperatures during the full glacial period were approximately 7°C cooler than present mean annual temperatures. Nivation is a term introduced by Matthes (1900) to denote the erosive effects associated with an immobile and patchy snow cover. Dohrenwend (1984) used the term nivation "to include the combined action of various periglacial weathering and erosion processes that are focused or intensified near the edges of long-

lived snow patches. It should be noted that the nivation threshold altitude during full glacial times rose from north to south at 190 m per degree of latitude, strongly suggesting significant climatic control of these landforms.

Packrat Middens

One of the principal objectives of this investigation is to integrate a model of geothermal fluid genesis with evolving models and concepts dealing with the paleohydrology of the Great Basin during the Wisconsin age. In the Great Basin, paleohydrology and paleoclimatology are invariably related to the distribution, extent, and fluctuations of pluvial lakes that occupied closed basins in Nevada, Utah and parts of Oregon and California. This will be discussed in detail in a Chapter 7.

The earliest efforts in determining the paleoclimate of the Great Basin were based on classical stratigraphy, sedimentology, and geomorphology of lake basins. The latest models are based on the analysis of plant material obtained from the preserved remains of packrat or woodrat nests: middens. Middens are biological accumulation of plants and other organic material collected by the species *Neotoma*, also known commonly as the woodrat, packrat and trade-rat.

Wells and Jorgenson (1964) first introduced the science of paleonidology as a new tool for the study of late Quaternary environments. Spaulding (1985) defined paleonidology as "the reconstruction of past vegetation from plant remains preserved in

middens deposited in rock cavities". Spaulding (1985) provides a complete description of midden composition, construction, preservation and morphology. The following is paraphrased from his description.

Currently active middens are loose debris piles, while ancient middens are usually masses solidified by dehydrated urine. Middens often serve as urination points, and cementation is caused by saturation with urine which, upon drying, encases the debris in a cohesive, crystalline mass.....indurated packrat middens are self sealing, with typically convoluted rinds that vary in thickness from a few millimeters to several centimeters. Middens must be sheltered from rain in a cave or beneath a rock overhang to be preserved....Some are more than 50,000 years old and are among the oldest fossils preserved by mummification.

The earliest efforts document the utility of packrat midden analysis for Quaternary studies. Several factors account for the success of packrat rat midden analysis over classical geologic investigations for certain types of Quaternary studies. All *Neotoma* species, for example, share the acquisitive behavior trait and construct houses and dens from collected plant debris from a limited foraging radius, estimated to be approximately 30 m, but certainly no more than 100 m. In addition to plant debris, middens contain fecal pellets, pollen, and skeletal remains. All the materials are tightly cemented by dried and crystallize urine, which is sometimes colloquially known as "amberat". The variety of fossil preserved material, the focused location and elevation, and the continuity (some provide continuous documentation over a period of 50,000 years), have allowed workers to accurately reconstruct paleoenvironments and paleoclimates throughout the southwestern United States.

Wells and Jorgenson (1964) working at the Nevada Test Site in southern Nevada, identified pinyon, curl-leaf mahogany, and Utah juniper in packrat middens where these higher-elevation species no longer flourish. This lead them to speculate that, between 40,000 and 7,800 years BP, the climate was colder and perhaps more moist than at present. Wells (1983) demonstrated on the basis of midden analysis, that during the late Wisconsin age, the Great Basin was dominated by subalpine forests of Bristlecone Pine and Juniper extending down to an elevation of approximately 1,660 m. Additional analyses showed that the present day desert vegetation replaced the pinyon juniper woodlands beginning 12,000 to 8,000 years BP.

Thompson and Mead (1981) confirmed earlier concepts of a widespread subalpine forest in the southern portion of the Great Basin between 25,000 and 11,000 years BP. Midden analysis indicated that the forest habitats were isolated islands during a climate of cool summer temperatures. They suggested that the widespread pluvial lakes were the result of a combination of the reduced summer temperatures and lower evaporation rates.

Spaulding (1985) documented major climatic changes in south-central Nevada over the last 45,000 years on the basis of packrat-midden analysis. Again, the macrofossil assemblage indicated a widespread juniper (high-elevation species) flourished throughout the region between 45,000 and 10,000 years BP. Webb (1986) completed a compilation of the distribution of packrat midden analyses and showed that the frequency of dates peaked at 11,000

years BP. The spatial distribution revealed a concentration of research activities in selected areas, including the Nevada Test Site and the Colorado River area.

The deuterium content of organic matter is known to vary systematically on the basis of origin of the sample (Schiegl and Vogel, 1969). Paleoclimatic reconstruction has been completed for the interval 9,500 to 22,000 years BP using ancient wood cellulose (Yapp and Epstein, 1977). Siegel (1983) was one of the first to take the midden technique a step further, extracting cellulose from plant debris contained within packrat middens collected in the Snake Range, eastern Nevada. The radiocarbon age of the midden material ranged from 34,000 years BP to present, and he was able to identify a significant deuterium depletion between 13,000 and 34,000 years BP (Figure 6.3). This time frame corresponds to the cool, wet period that previous authors (Wells and Jorgenson, 1964; Wells, 1983; Thompson and Mead, 1981; and Spaulding, 1981) have identified from 40,000 to 10,000 years BP on the basis of plant species alone. Due to the possibility of significant intramidden variation in deuterium content, Siegel (1983) suggested restricting the sampling and analysis to one type of plant material (ie. twigs of Utah juniper).

In the course of this study, 25 samples of Utah juniper from 5 sites in northern and central Nevada were obtained and submitted for deuterium analysis. Figure 6.4 shows the relationship between radiocarbon age and deuterium content of middens from the Painted Hills, located just north of Reno, Nevada. The Painted Hills site

Figure 6.3: Deuterium vs. carbon-14 age of packrat middens from the Snake Range (Siegel, 1983).

Figure 6.4: Deuterium vs. carbon-14 age of packrat middens from the Painted Hills location north of Reno, Nevada.

was chosen for detailed study due to its long history of occupation and the large number of analyses available. Like the Siegel (1983) study, the data reveal a significant deuterium depletion extending from approximately 12,000 to 25,000 years BP. The two data sets are shown together in Figure 6.5. Although both show a period of deuterium depletion, the age of onset is different (13,500 vs. 11,500 years BP).

Several factors may be considered to account for the 2,000 year difference in the deuterium depletion onset. The error bars for radiocarbon ages, not shown in the diagram, may account for several hundred years, but cannot explain the entire discrepancy. The difference in elevation, latitude and longitude of the study locations may indicate a different timing for the cessation of the

Figure 6.5: Deuterium vs. carbon-14 age of packrat middens from the Snake Range (Siegel, 1983) and the Painted Hills (this study). Note the similar patterns of depletion and enrichment.

pluvial climate across the region. The aspect of the slope on which the middens are located may also be important. South and west facing slopes will receive greater solar insolation which may be reflected in the isotopes of the collected material. A final important factor, to be discussed in detail in Chapter 7, is the location of the two sites relative to Quaternary age pluvial lakes. The Snake Range is located on the Nevada-Utah border, adjacent to the southwest arm of pluvial Lake Bonneville (Snyder and others, 1964). Most of the midden material collected during this investigation was derived from the Lake Lahontan Basin, with the Painted hills midden located only a few hundred meters away from the Lake Lahontan highstand shoreline.

Pluvial Lakes in the Great Basin

In order to address the timing of deuterium depletion, it is necessary to review the literature, concepts and models pertaining to pluvial lakes in the Great Basin. During the Pleistocene, the Great Basin was occupied by over 100 pluvial lakes (Snyder et al., 1964), at their highstands inundating over 20 percent of the land surface (Houghton et al., 1975). Snyder et al. (1964) were among the first modern authors to map the aerial extent of Lake Lahontan and Lake Bonneville. They concluded that the highest lake stages occurred when the combination of temperature, precipitation, and evaporation produced a maximum inflow and minimum rate of evaporative loss. Broecker and Kaufman (1965) produced a similar map (Figure 6.6) and identified the timing of four highstands in the Late Wisconsin: 17,000, 14,500, 12,000 and 9,500 years BP. Mifflin

and Wheat (1979) determined on the basis of geomorphology that the last high stand of Lake Lahontan was between 1,320 and 1,340 meters above sea level, and that this time period experienced an average temperature almost 3 degrees Centigrade cooler than the present.

Benson (1978) identified two high stands at 1,330 m, the first occurred between 25,000 and 22,000 years BP and the most recent extended from 13,500 to 11,000 years BP. Benson (1981) also determined that percentage of cloud cover was an important factor in the filling and evaporative loss equation for pluvial lakes. In 1987, Benson and Thompson identified the timing of a high stand (1,330 m

Figure 6.6: Maximum extent of major pluvial lakes in the Great Basin, from Benson et al. (In Press).

at 13,500 years BP) and a low stand (1,170 m at 5,000 years BP) for pluvial Lake Lahontan.

In the Walker Lake Basin, Benson (1988) identified the timing of two high stands (14,000 and 12,500 years BP) and three low stands (21,000 to 15,000, 4,700, and 2,600 years BP), based on interpretation of cored lake sediments. The high stands correlate with most data from the northern portions of Lake Lahontan. Davis (1983) determined that Lake Lahontan attained a level of 1,256 to 1,260 m approximately 23,400 years BP in the Trego Hot springs area near Gerlach, Nevada. The data are from analysis of volcanic tephra and suggest that the level of the lake was rising at this time.

Williams and Bedinger (1983) mapped the probable distribution of Pleistocene marshes (as well as lakes) in the Great Basin. Their work is based on the assumption that present-day closed basins with a groundwater table within 10 meters of the surface probably contained a marsh during the Pleistocene. The distribution of Pleistocene marshes suggests that the storm systems that supplied the northern Great Basin with sufficient precipitation to produce Lakes Lahontan and Bonneville, were sufficiently dispersed to maintain saturation of the groundwater systems in the southern Great Basin as well. At the Nevada Test Site, Winograd and Doty (1980) determined that groundwater levels between 40,000 and 10,000 years BP may have been 50 m higher than the present.

Thompson et al. (1986) show that the last major high stand of Lake Lahontan terminated before 12,070 years BP and that major

fluctuations were nearly synchronous with Lake Russell (Mono Lake Basin, California) and Lake Bonneville. The evidence suggests a gradual rise in lake levels between 24,000 and 16,500 years BP, a rapid rise (1,270 to 1,330 m) from 16,500 to 13,500 years BP, followed by a 1,000 year period of little variation. This quiescent stage was followed by a rapid drop from 1,330 to 1,170 m in less than 500 years. The most recent lake-level model (Benson et al., In press) incorporates lake level changes for Lakes Lahontan, Bonneville, Russell and Searles during the last 35,000 years. Due to the complexity of the model, it is impractical to fully describe the fluctuation patterns. The curve for Bonneville and Lahontan Lake level fluctuations is shown in Figure 6.7.

Figure 6.7: Lake level chronology of pluvial Lake Lahontan and Lake Bonneville, from Benson et al. (In Press).

CHAPTER 7. INTERPRETATION

REGIONAL ISOTOPIC DIFFERENCES

It has been shown that isotopic data for fluids from various sources within a particular study area are often represented graphically by plotting oxygen-18 vs. deuterium. Typical geothermal fluids in the Great Basin are isotopically depleted with respect to non-thermal fluids discharging at the same elevation. This "difference" occurs throughout the Great Basin and is relative to the absolute values of the isotopic composition of the two fluid types. Figure 7.1 is an example showing the relationship between thermal and non-thermal fluids from the Beowawe area in north-central Nevada. Figure 7.2 shows similar data from the Caliente area in southern Nevada. The relative difference of deuterium in fluids from the northern portion of the study area is approximately equal to the relative difference between the two fluid types in the south. But the regional significance of this relative difference is difficult to assess using data in this format.

In order to examine the systematic regional variations in the isotopic differences between thermal and non-thermal fluids, a technique was developed to "subtract" the interpolated isotope values of one data set from the interpolated values of the other. Figure 7.3 illustrates the regional deuterium differences between thermal wells and non-thermal springs throughout the Great Basin. In this case, a temperature greater than or equal to 80°C was used for the geothermal proxy and temperatures from springs less than or equal to 20°C was used as a modern precipitation proxy. In Figure

7.3, positive values (contours 0 and greater) indicate the areas where cold springs (modern meteoric water proxy) are isotopically more depleted than local thermal wells. Negative values (contours less than 0) show the areas where thermal fluids that are more depleted than the cold water proxy.

As expected, thermal fluids in the northern and western portions of the Great Basin are depleted in deuterium content by 8 to 12 permil, when compared to the modern proxy. The data in the southeast are spurious because there are no thermal wells in that area with temperatures in excess of 80°C. In the north central portion of the study area, in the vicinity of the Carson Sink, the data suggest that thermal fluids are more "enriched", instead of more depleted, than the modern proxy.

Using springs (rather than wells) greater than 80°C as the geothermal proxy produces a similar configuration, but the contours are somewhat offset (Figure 7.4). In addition, Figure 7.3 demonstrates that thermal well water is generally isotopically more depleted than thermal spring water. Geothermal wells are cased with steel pipe which essentially isolates the reservoir and precludes contamination with near surface groundwater.

If oxygen-18 is contoured rather than deuterium, and the two data sets are subtracted from one another (Figure 7.5), (springs >80°C minus springs <20°C), a different configuration is achieved. The diagram shows that thermal fluids are not more isotopically depleted than the modern proxy. This seeming contradiction may be

Figure 7.3: Deuterium permil difference for thermal wells minus non-thermal springs. The thermal wells are more isotopically depleted than the non-thermal springs, indicated by the negative (solid) contour lines. Dashed contour lines define areas where thermal wells are relatively enriched.

Figure 7.4: Deuterium permil difference for thermal springs minus non-thermal springs. Thermal springs are also more isotopically depleted than the non-thermal springs, except in areas defined by positive (dashed) contour lines.

explained on the basis of the behavior of oxygen isotopes at high temperature.

Figure 7.1, which shows a plot of oxygen-18 vs. deuterium for the Beowawe, Nevada, area, demonstrates that although the relative deuterium difference is approximately 12 permil, the oxygen-18 difference is less than 1 permil. Craig (1963) reported that rock-water interactions at high temperature yield a fluid that is more enriched in oxygen-18 than the fluid's initial concentration. This "oxygen shift" is characteristic of high-temperature geothermal

Figure 7.5: Oxygen-18 permil difference for thermal springs minus non-thermal springs. The non-thermal springs are everywhere more isotopically depleted than the thermal springs, opposite of the occurrence with plots of deuterium concentrations.

fields throughout the world, including Steamboat Springs, Nevada, Salton Sea, California, Yellowstone National Park, Wyoming, and Lardarello, Italy (Figure 7.6). Oxygen-18 therefore will not be used to delineate recharge areas in this investigation.

The significance of the deuterium difference for thermal wells and springs may be further illustrated by noting the relationship between "positive" isotope regions and the location of pluvial lakes that occupied large portions of the Great Basin during the

Figure 7.6: Examples of 'oxygen-shift' from geothermal fields throughout the world. T's represent thermal fluids, while N's represent non-thermal waters.

Pleistocene. Figure 7.7 shows the outline of the highstands of Lake Lahontan and Lake Bonneville superimposed over the previously discussed deuterium enriched areas. The diagram shows that all of the "deuterium enriched" areas are co-located with large pluvial lakes and suggests that the lakes participated in geothermal fluid recharge during the Pleistocene.

The areas occupied by major pluvial lakes also correspond to the highest concentrations of sodium chloride-type fluids in the

Figure 7.7: Areas where non-thermal springs are isotopically more depleted than thermal water in relation to major pluvial lakes. Long dash lines indicate thermal wells and short dash lines indicate thermal springs. Contours represent permil deuterium difference from Figures 7.3 and 7.4.

Great Basin. In addition to isotopic enrichment, these fluids are characterized by geochemical enrichment. These two processes are the result of evaporation and subsequent concentration of isotopes and ions within the pluvial lakes.

It is important to examine the nature of both depleted and enriched fluids as a potential source of recharge. To accomplish that, it will be necessary to examine the paleoclimate record to determine the relationship between pluvial lake highstands and isotopic depletion.

PLUVIAL LAKE HIGH STANDS AND THE ISOTOPIC RECORD

The previous section illustrated the utility of an interdisciplinary approach to the resolution of geothermal fluid recharge by correlating the location of pluvial lakes with anomalous areas of isotopically enriched thermal fluids. This section discusses the timing of fluid recharge in terms of fluid age, isotopic composition, pluvial lake highstands, and paleoclimate proxies.

Figure 7.8 shows a comparison between the Bonneville Lake level fluctuations and the variation in the deuterium content of plant material from packrat middens from the Snake Range on the Utah-Nevada border (Siegel, 1983), for the last 40,000 years. Figure 7.9 is a similar plot using Lahontan lake level fluctuations and the packrat midden data set from the Painted Hills, western Nevada. Both figures illustrate an important concept.

From 40,000 and 10,000 years BP, highstands of Lake Bonneville and Lake Lahontan were synchronous with significant depletions in the deuterium content of terrestrial plants growing in the Great Basin. However, the onset of deuterium depletion appears to be a sub-regional event that occurred at different times throughout Nevada and Utah. A minimum value for deuterium is recorded in the Bonneville Basin at approximately 17,500 years BP. The minimum deuterium value for the Lahontan Basin corresponds to an age of 12,500 years BP.

Several factors must be considered to account for this apparent discrepancy. The error bars for radiocarbon ages, not shown in the diagrams, will only account for several hundred years of the discrepancy. Differences in elevation, latitude and longitude for the two sites may also account for some of the discrepancy. The aspect of the slope that contains the midden is also important. South and west facing slopes receive a greater percentage of solar insolation, which results in warmer temperatures, increased evaporation rates, and isotopic enrichment.

Another factor that could account for the different timing is the basin topography of Lake Bonneville and Lake Lahontan. Benson et al. (in press) note that Lake Lahontan's rise was not gradual (as was Bonneville's) due to topographic restraints and intrabasin spill-over. The author's suggest that parameters such as lake size, volume, and level may have been affected by a combination of factors including a change in the direction of the jet stream over

Figure 7.8: Normalized deuterium from Snake Range packrat middens and normalized Bonneville lake level variations. Siegel midden study location is adjacent to the southwest arm of Lake Bonneville.

Figure 7.9: Normalized deuterium from Painted Hills midden samples compared to Lahontan lake levels. Painted Hills location is within a few hundred meters of the Lake Lahontan highstand shoreline.

the Great Basin, and a lake-atmosphere feedback mechanism that produced a microclimate in the vicinity of the larger lakes.

Thus, the middens and lake levels probably record the effects of at least two Great Basin microclimates imprinted on the global climatic events of the late Pleistocene. The data from both global and Great Basin studies have demonstrated a distinct δD depletion that lasted from approximately 40,000 years to 10,000 years BP. What is the relationship between the isotopic depletion, geothermal fluids, and recharge?

To address that relationship, it is necessary to review the available data base on a regional scale. Of particular importance is the relationship between the fluid's radiocarbon age and its deuterium value. Data from ten sites in northern and southern Nevada have been incorporated into this analysis.

As discussed previously, an extensive data base of radiocarbon ages has been created through work in the vicinity of the Nevada Test Site and Yucca Mountain, and is listed in Table 7.1. Figure 7.10 shows a series of curves from several of these southern Nevada groundwater investigations (Grove et al., 1969; Claassen, 1985; Claassen, 1986; White and Chuma, 1987; and Lyles and Hess, 1988). The diagram shows the characteristic deuterium depletion for fluids with an estimated age in excess of 10,000 radiocarbon years. The fluid temperatures range from 30° to 40°C and the range of deuterium values, -92 permil to -114 permil is consistent with groundwaters in southern Nevada.

Table 7.1 - Radiocarbon age data from the literature search.

SAMPLE	AGE	DEUT	LOC	SAMPLE	AGE	DEUT	LOC
SAMPLES FROM SOUTHERN NEVADA							
<u>White and Chuma, 1987</u>				FAIRBANK	17869	-106.0	AD
W1	15252	-112.5	OV	POINT RKS	15611	-110.0	AD
U20a2	12525	-114.0	FC	IS SP	17498	-104.0	LV
W28	9394	-103.0	AD	WELL2	17259		YF
8ABB	7968	-99.5	AD	U20a2	16475	-114.0	PM
8DCC	8914	-103.0	AD	WELL5B	19733	-110.0	YF
J13	7888	-97.5	FC	WELL8	4397	-104.0	PM
J12	7269	-97.5	FC	CORN CREEK	5762	-105.0	LV
W13	4380	-108.0	OV				
W6	3872	-102.0	OV	<u>Claassen, 1985</u>			
W2	2855	-110.5	OV	FAIRBANK	28000	-103.0	AD
UE29a2	3792	-93.0	FC	C27	15651	-105.0	AD
UE29a2	3317	-93.5	FC	C16	13000	-104.0	AD
W15	0	-108.0	OV	8DCC	12828	-104.0	AD
W17	0	-104.0	OV	VH1	16000	-108.0	CF
W3	0	-102.0	OV	ASHTREE SP	11000	-102.0	AD
W5	0	-108.0	OV	C25	10685	-102.0	AD
				C3	11000	-102.0	AD
<u>Lyles and Hess, 1988</u>				C23	10952	-103.0	AD
MARTIN	28000	-100.0	LV	C15	8000	-104.0	AD
ISPRIS	20000	-102.0	LV	C17	8000	-105.0	AD
POINTB	18800	-98.0	LV	C4	8926	-103.0	AD
CC WL	18300	-95.0	LV	C13	9000	-102.0	AD
IS AFB	17100	-96.0	LV	H1	13100	-102.0	FC
IS SP	17600	-93.0	LV	C11	7000	-101.0	AD
HOLL	12600	-98.0	LV	8ABB	9158	-99.5	AD
TULE	11900	-99.0	LV	C8	8580	-95.0	AD
CC SP	13300	-96.0	LV	C10	6604	-97.5	AD
MIFLIN	11300	-100.0	LV	WELL8	11000	-104.0	PM
STEW	6200	-100.0	SM	C21	7375	-99.0	AD
GRAPE	4100	-91.0	SM	C18	5000	-102.0	AD
CORT	2500	-96.0	LV	C9	5000	-102.0	AD
SANDST	4600	-90.0	SM	UE25b1	10000	-101.0	FC
ADAMS	2600	-98.0	LV	C1	7455	-97.5	FC
PAHUTE	2000	-98.0	LV	C47	5038	-102.0	AD
GILBER	1500	-98.0	SM	C2	6960	-97.5	FC
RED SP	2900	-93.0	SM				
WILLOW	100	-98.0	LV	<u>Boughton, 1986</u>			
BIGTIM	0	-93.0	AD	WELLC1	30400	-102.0	CF
WIREGR	0	-94.0	SH	WELLA	21400	-107.0	CF
MAINT	0	-96.0	SM	Ue15d	13800	-99.0	CF
DEER	0	-98.0	SM	Ue16d	19600	-106.0	CF
MTCHAR	0	-96.0	SM	U19c	21500	-109.0	PM
				WELL2	18500	-102.0	CF
<u>Grove et. al., 1969</u>				WELL4	12600	-94.0	CF
WELLC1	30168	-114.0	YF	WELL8	10900	-102.0	PM
WELL1	22590	-107.0	LV	J13	6590	-96.0	FC

Table 7.1, continued

SAMPLE	AGE	DEUT	LOC	SAMPLE	AGE	DEUT	LOC
<u>Benson, 1987</u>				<u>Jacobsen et. al., 1983</u>			
	16000	-105.0	YM	HYDER HS	24864	-130.0	DV
	12800	-101.0	YM	SOU HS	26995	-132.0	DV
USWH4	15000	-104.0	YM	MCCOY HS	17125	-122.0	DV
USWVH1	15500	-108.0	CF	ART1	21691	-134.0	DV
	14000	-107.0	YM	DIXIE HS	4375	-120.0	DV
	12500	-102.0	YM	KYLE HS	12042	-130.0	DV
	12500	-103.0	YM	MIDDLEGATE	5306		DV
USWH6	12000	-106.0	YM	WHITE	430		DV
	11500	-100.0	YM	PARK	-139		DV
Ue25b1	14200	-101.0	YM	<u>Glancy, 1986</u>			
USWH5	23500	-102.0	YM	FL	8800	-112.5	FL
UE25b1	12000	-99.5	YM	FL	3500	-110.5	FL
USWH5	12000	-102.0	YM	FL	4400	-114.0	FL
USWG4	11000	-103.0	YM	FL	4200	-112.5	FL
USWH1	11850	-101.0	YM	FL	1100	-110.5	FL
J13	7000	-97.5	FC	FL	1500	-108.0	FL
J12	7000	-97.5	FC	FL	1500	-107.5	FL
UE29a2	5000	-93.0	FC	FL	1400	-114.0	FL
UE29a2	5000	-93.5	FC	FL	0	-110.0	FL
SAMPLES FROM NORTHERN NEVADA				FL	0	-111.0	FL
<u>Flynn and Ghusn, 1984</u>				<u>Szecsody, 1982</u>			
STEAMBOAT	43300	-115.3	SB	CARSON	13952	-128.0	EV
NEWBURG	28030	-128.0	MO	EV48	2617	-111.0	EV
WARREN EST	13680	-128.9	MO	PRISON	0	-111.0	EV
GADDA	9530	-122.9	MO	EV52	212	-110.0	EV
MILES	8200	-130.5	MO	EV10	0	-126.0	EV
DESJARDIN	3140	-122.4	MO	GRAHAM	0	-108.0	EV
MCKAY	1680	-108.8	MO	<u>Nehring, 1980</u>			
MOANA LANE	910	-113.2	MO	SB26	44491	-115.8	SB
ZOLEZZI L.	0	-93.9	MO	SB8	38323	-115.9	SB
				SB5	36495	-114.9	SB
				SBNR27	33981	-115.6	SB

Key to sample locations

SOUTHERN NEVADA

LV - Las Vegas Valley
 AD - Amargosa Desert
 FC - Fortymile Canyon
 CF - Crater Flat
 OV - Oasis Valley
 PM - Pahute Mesa
 SM - Spring Mountains
 YF - Yucca Flat
 YM - Yucca Mountain

NORTHERN NEVADA

MO - Moana District
 FL - Fallon Area
 SB - Steamboat Springs
 EV - Carson-Eagle Valley
 DV - Dixie Valley

Figure 7.11 shows a similar set of curves using data from various groundwater investigations in northern Nevada. Data analyses acquired during this study are listed in Table 3.3, other radiocarbon age data for northern Nevada, including moderate temperature fluids (Glancy, 1986), and thermal fluids (Flynn and Ghusn, 1984; Bell et al. 1980; Szecsody, 1982; Jacobson et al. 1983), are listed in Table 7.1. Fluid temperatures range from 30° to 98°C, and the total range of δD values is consistent with sites in northern Nevada. Once again, the curve shows the characteristic deuterium depletion between 35,000 and 10,000 years BP, in this case radiocarbon years.

Various authors have expressed uncertainty regarding the reliability of radiocarbon age dating of groundwater (Grove et al., 1969; White and Chuma, 1987). In particular, atmospheric and soil carbon corrections are estimated on the basis of modern values (see Appendix 2 for a more detailed discussion). There appears to be more confidence in radiocarbon data from non-aqueous sources such as wood, bone, charcoal, and peat.

Figure 7.12 shows the deuterium vs. radiocarbon age relationship between northern Nevada thermal fluids and the two packrat midden data sets. There is a clear correlation between the onset of deuterium depletion at approximately 30,000 years BP for all three curves. Maximum depletion for thermal fluids occurs at 25,000 years, with a gradual increase to modern values over a 15,000 year period. The shape of the curve is partially reflected in the midden data, but neither curve illustrates the magnitude of

Figure 7.10: Deuterium vs. carbon-14 age from several adjacent southern Nevada study areas. Y = Yucca mountain, F = Fortymile Canyon, and A = Amargosa Desert. Note trend from modern isotopic enrichment to ancient depletion.

Figure 7.11: Deuterium vs. carbon-14 age from several northern Nevada study areas. S = Steamboat Springs, C = Carson City, D = Dixie Valley, M = Moana District, and F = Fallon.

depletion. Although the middens reflect the change from enriched to depleted isotopes, thermal fluids are considerably more depleted. Possibly, vadose-zone water absorbed by the plants is subject to more evaporation, resulting in slightly more enriched isotopes.

A similar correlation can be made between pluvial lake high stands and the deuterium depletion record for thermal fluids. (Figure 7.13). These data are shown on a normalized scale for comparative evaluation. The onset of deuterium depletion is shown to correlate with the initiation of lake level rise in both the Lahontan and Bonneville basins. The deuterium depleted precipitation that established the lakes also contributed fluids to the geothermal systems.

FLUID TEMPERATURES AND RADIOCARBON AGE

There does not appear to be a systematic correlation between fluid age and temperature. This suggests that flow paths, and depths of circulation are unique for each geothermal system. In order to address this issue, data from seven geothermal systems in Nevada were evaluated in terms of estimated fluid age and regional geothermal gradients.

Average temperature gradients range from 40° to 60°C per km depth throughout the Great Basin (Garside and Schilling, 1979). Temperature gradients from shallow temperature exploration holes are misleading and may be as high as 600°C per km in the vicinity of a near-surface geothermal resource. On the basis of fluid age,

Figure 7.12: Deuterium vs. carbon-14 age of packrat middens, and thermal waters greater than 80°C from various locations in northern Nevada.

Figure 7.13: Deuterium vs. carbon-14 age of thermal waters greater than 90°C, compared to the level chronology of Lake Lahontan and Lake Bonneville. Note how the trends are inversely related.

and an average geothermal gradient of 40°C per km, the average minimum depth to circulation may be estimated (Table 7.2).

It should be noted that these calculations are based on a simple linear relationship between fluid age and temperature gradient. On this basis, flow rates for a round-trip range from 0.1 to 0.2 m per year. The variation in depth of circulation and fluid age suggest that the travel time curve is not symmetrical. These depth estimates are consistent with the estimated depths of penetration of range-bounding faults (10 to 15 km) and estimates of depth of circulation of fluids in fault zones (3 to 5 km) of Zoback and Anderson (1983).

Table 7.2: Calculated average minimum depth of circulation for several geothermal systems in northern Nevada.

LOCATION	PRODUCTION ZONE DEPTH (km)	FLUID AGE (ka)	ESTIMATED TEMP. (°C)	MINIMUM DEPTH OF CIRCULATION (km)
Steamboat	1.0	43	204	5.1
Moana	0.3	28	151	3.7
Warm Spring	---	42	100	2.5
Elko	0.3	34	120	3.0
Dixie Valley	3.0	30	260*	6.5
Desert Peak	1.8	44	200	5.0
Beowawe	3.0	23	200	5.0

(* estimated from unpublished reports, all others are Quartz Geothermometer, with no steam loss.)

TIME VARIANT ANALYSIS OF ISOTOPIC DATA

This portion of the study was designed to identify systematic trends in isotopic values of geothermal power plant production wells. Several problems became apparent in the course of the analysis.

The overall data set contains duplicate fluid analyses from several locations. The type of duplication ranges from two samples taken 10 to 20 years apart and collected during separate studies, to a systematic monthly sampling program for a specific site.

The marked and consistent difference between samples analyzed in the 1970's and those analyzed in the 1980's raised questions regarding the overall laboratory reproducibility. For example, samples from Winograd and Pearson (1976) and Winograd and Friedman (1972) are consistently 5 to 10 permil more depleted than recent studies (Kirk, 1987; Lyles and Hess, 1988, This Study, 1989) of the same sites (Table 7.3). It is also difficult to determine exactly which fluid source was sampled, and the sampling procedure used. Fluid fractionation can occur at any step of the sampling procedure, and good judgement must be exercised in the sampling program to minimize errors.

A systematic time-variant analysis by Lyles (1985) in the Reno, Nevada, area demonstrated the yearly variation possible at a single site (Figure 7.14). The five thermal sources studied were sampled either monthly (Bowers Mansion) or bi-monthly (all others) for one full year (Table 7.4). The fluids showed an average yearly range in deuterium content of 13‰ . Recent meteoric water is the

most likely source of the seasonal variation in the thermal fluid. Lyles correlated precipitation events and isotopic response on a linear basis for three of the sites. Different flowpaths for each system would account for the varied timing of isotopic change. Time-variant analysis of isotopes conducted over a period of years must consider seasonal changes and the weather for each year.

Table 7.3: Comparison of historical deuterium analyses from southern Nevada with recent studies.

SAMPLE LOCATION	REF	LONG	LAT	DEUT	MO YR
Hot Creek Spring	KRK	115.15	38.38	-104	87
Hot Creek Spring	WNF	115.15	38.38	-124	3 70
Hiko Spring	KRK	115.21	37.59	-107	87
Hiko Spring	WNF	115.21	37.59	-113	3 70
Hiko Spring	WNF	115.21	37.59	-113	1 69
Hiko Spring	WNF	115.21	37.59	-114	8 68
Crystal Spring	KRK	115.32	37.53	-109	87
Crystal Spring	WNF	115.32	37.53	-112	3 70
Crystal Spring	WNF	115.32	37.53	-112	8 68
Crystal Spring	WNF	115.32	37.53	-113	1 69
Ash Spring GF-19	GFG	115.19	37.46	-111	2 89
Ash Spring	KRK	115.19	37.46	-108	87
Ash Spring	WNF	115.19	37.46	-115	3 70
Ash Spring	WNF	115.19	37.46	-112	1 69
Ash Spring	WNF	115.19	37.46	-110	8 68
Indian Springs	LYL	115.66	36.56	-93	6 86
Indian Springs	WNF	115.66	36.56	-104	3 70
Indian Springs	WNF	115.66	36.56	-103	1 69
Indian Springs	WNF	115.66	36.56	-101	8 68
Nevares Spring GF-16	GFG	116.79	36.51	-105	2 89
Nevares Spring	WNF	116.79	36.51	-109	3 70
Nevares Spring	WNF	116.79	36.51	-111	8 68

WNF - Winograd and Friedman, 1972

WNP - Winograd and Pearson, 1976

GFG - This Study

KRK - Kirk, 1987

LYL - Lyles and Hess, 1988

Figure 7.14: Time-variant isotopic analysis of several thermal sources in the Reno-Carson City area (Lyles, 1985). B - Bowers Mansion, SB - Steamboat F- Farad, W - Walleys, ST - Saratoga.

Table 7.4: Time-variant deuterium analysis study on several Reno, Nevada area warm springs, Lyles (1985).

Saratoga Hot Spring DEUT mo yr	Walleys Hot Spring DEUT mo yr	Bowers Mansion DEUT mo yr
-126 9 83	-122 9 83	-109 9 83
-124 11 83	-115 11 83	-105 10 83
-128 1 84	-113 1 84	-104 11 83
-123 3 84	-109 3 84	-105 12 83
-119 5 84	-111 5 84	-108 1 84
-109 7 84	-109 7 84	-106 2 84
Steamboat Springs DEUT mo yr	Farad Warm Spring DEUT mo yr	-103 3 84
-112 10 83	-108 9 83	-102 4 84
-109 12 83	-115 11 83	-105 5 84
-106 2 84	-106 1 84	-101 6 84
-105 4 84	-107 3 84	-104 7 84
-109 6 84	-102 5 84	-105 8 84
-120 8 84	-106 7 84	

Less variation is expected in a deep reservoir, as seasonal changes have little effect. Data from three Chevron Resources geothermal power plants do, however, demonstrate a distinct non-seasonal depletion in deuterium (Table 7.5, Figure 7.15). The question of laboratory reproducibility must again be addressed here. Data from 1987 and 1988 was provided by Chevron Resources, Inc., and was analyzed by a different laboratory than samples from this study. The timing of the January, 1989, depletion appears to

Table 7.5: Chevron power plant production well isotope data. 1987 and 1988 data from Chevron Resources, Inc., archives. 1988 data from this study. Letters in column 2 relate to figures 7.15 and 7.16.

GF #	#	SAMPLE	DEUT	OX18	T (c)	DATE mo yr
10	A	Beowawe Ginn 1-13	-133.9	-15.67	142	5 87
	B	Beowawe Ginn 1-13	-130.8	-15.72	142	5 87
	C	Beowawe Ginn 2-13	-132.8	-15.54	151	5 87
	D	Beowawe Ginn 2-13	-130.9	-15.59	151	5 87
	E	Beowawe Ginn 2-13	-125.6	-15.35	154	10 88
	F	Beowawe Ginn 1-13	-123.8	-15.26	151	10 88
	G	Beowawe Ginn 1-13	-131.5	-16.20	150	1 89
11	H	Beowawe Ginn 2-13	-132.6	-15.90	150	1 89
34	I	Beowawe Ginn 1-13	-131.4	-15.90	150	4 89
46	J	Beowawe Ginn 2-13	-131.1	-16.10	150	10 89
9	A	Desert Peak 67-21	-110.6	-12.30	162	5 87
	B	Desert Peak 86-21	-109.7	-12.13	163	5 87
	C	Desert Peak 86-21	-109.3	-11.97	162	10 88
	D	Desert Peak 67-21	-108.7	-11.58	162	10 88
	E	Desert Peak 86-21	-114.3	-12.10	163	1 89
	F	Desert Peak 67-21	-114.9	-12.30	163	1 89
	G	Desert Peak 67-21	-118.1	-12.10	163	4 89
7	A	Soda Lake 84-33	-101.4	-10.98	185	10 88
	B	Soda Lake 84-33	-104.6	-10.80	183	1 89
	C	Soda Lake 84-33	-103.5	-11.20	183	4 89
	D	Soda Lake 84-33	-105.9	-11.17	183	10 89

coincide with the laboratory change. However, samples from Desert Peak suggest the phenomena is not a laboratory artifact, as the depletion is recorded in consecutive samples from this study. Unfortunately, the plant was shut down for maintenance and not included in the October, 1989, sampling series.

Oxygen-18 (Figure 7.16) undergoes a similar, but much smaller depletion. The percent apparent depletion for oxygen-18 and deuterium concentrations is similar.

The probability that fluids from all three power plants, which tap different reservoirs, experience a simultaneous isotopic depletion is remote. Laboratory or sampling procedure inconsistency is more likely responsible for the apparent depletion. Continued monitoring on a systematic basis under controlled conditions may be effective in detecting early changes in fluid reservoir conditions.

Figure 7.15: Deuterium vs. time for Chevron power plant production wells, from Chevron Resources archives (1987, 1988 data) and current study (1989 data). Letters on curves refer to data listed in Table 7.5.

Figure 7.16: Oxygen-18 vs. time for Chevron power plant production wells, from Chevron Resources archives (1987, 1988 data) and current study (1989 data). Letters on curves refer to data listed in Table 7.5.

CHAPTER 8. CONCEPTUAL MODEL OF GEOTHERMAL FLUID RECHARGE

Historically, geothermal recharge models have been based on site specific investigations that focused on current meteoric conditions. The advantage of this contemporary approach is the ability to develop detailed reservoir information that can be integrated into production schedules, determine well spacing, field life, and other engineering parameters. Difficulties arise when recharge models from one drainage basin or geothermal field do not corroborate data from an adjacent area (Welch, 1981; Higgins et al. 1985).

The classic form of the historical recharge model is illustrated in Figure 8.1. The model proposes that fluids are recharged by modern precipitation falling on adjacent mountain highlands. Infiltration is rapid and direct; fluids from specific elevations

Figure 8.1: Historical recharge model, which relied upon rapid recharge to circulate modern meteoric water great distances, laterally from the range-top source and vertically to and from the heat source or reservoir.

travel downward to the high-temperature zone, are heated, and return to the surface in the form of hot springs. Normal faults are identified as the principal conduits for upwelling thermal fluids. To account for isotopic variations, such as deuterium or oxygen-18 shifts, a diagram such as Figure 8.2 is used in conjunction with the conceptual model. This type of model was widely accepted, and was replicated in studies throughout the Great Basin.

If regional and temporal considerations are ignored, the historical model accurately reflects all of the physical and chemical requirements of recharge for a single system. Data presented in this report, however, have indicated that the isotopic content of geothermal systems in the Great Basin likely reflect the isotope geochemistry of precipitation in the Great Basin during the late Pleistocene, 40,000 to 10,000 years BP.

The new conceptual model proposed in this study consists of the following key elements:

1. the availability of significant volumes of water,
2. a regional deuterium depletion that varies systematically in concert with modern values,
3. an extended period of recharge,
4. a recharge mechanism that is consistent with the existing structural setting of numerous deeply cutting extensional faults (Figures 2.2 and 2.3),
5. and concurrence with thermal fluid isotope geochemistry.

Figure 8.2: Known trends in isotope ratios that have been incorporated into recharge models to explain compositional variations (modified from Bell et al., 1980)

Geothermal fluids from the Great Basin have apparent ages that range from 40,000 to 10,000 years BP. During this period, it is well documented that basins in parts of Nevada, southern Oregon, western Utah, and eastern California were inundated by more than 120 pluvial lakes. In addition, paleoclimate data has shown that precipitation during this time frame was isotopically depleted worldwide, relative to modern values. It has also been demonstrated that the isotopic composition of thermal fluids varies systematically throughout the Great Basin in a configuration that mimics modern precipitation, but at more depleted levels. These are the regional aspects of geothermal fluid recharge which must be factored into the final model.

The location of the Quaternary lakes played a significant role in the development of this model. A lake surrounded by a high water table, supplied by a non-homogeneous, anisotropic aquifer system with horizontal flow $\gg$ vertical flow, will experience groundwater inflow (seepage) through its base (Figure 8.3). The introduction of a high-conductivity zone (incorporating the aspects of permeability and porosity) will significantly alter the potentiometric field surrounding the lake (Fetter, 1981). This change in the hydraulic gradient will cause lake waters to leak through the most permeable portions of the lake bottom (Figure 8.4) and be drawn into the high-conductivity zone. This zone serves as a "drain" for waters located above, substantially increasing the component of vertical flow.

Figure 8.3: Flowpath for lake seepage and range run-off when inhomogeneous, anisotropic sub-surface conditions exist with horizontal flow $\gg$ vertical flow (Fetter, 1981). Lake will experience inflow through its base from all sides.

Figure 8.4: Suggested path for lake seepage and range run-off. Mixing occurs in the flooded fan. The fault serves as a conduit to recharge deep geothermal reservoirs with a slightly isotopically enriched fluid.

If we apply this scenario to the conditions that prevailed in the Great Basin during the Pleistocene, the following observations can be made.

1. Lake levels rose during the Pleistocene in response to lower mean-annual temperatures, and increased precipitation and cloud cover. Continental glaciation in the central part of North America pushed the jet stream several degrees latitude to the south, bringing "Seattle-like" weather to the Great Basin.

2. In most of the basins, runoff from precipitation and snow melt flooded alluvial fans with depleted meteoric fluids, some of which leaked into range-bounding faults beneath the alluvial fans. Concurrently, the lakes rose tens to hundreds of meters flooding the highly porous and permeable alluvial fans that border the basins. The most notable Lakes were Bonneville, Lahontan, and Russell. These lake waters were enriched in both isotopes and total dissolved solids (TDS) as the result of evaporation from the surface of the lakes. This is consistent with the interpretations of the high TDS thermal fluids near Brady's Hot Springs (Welch and Preissler, 1986). As water levels rose, the lake contributed isotopically enriched fluids to the groundwater system via the porous alluvial fans. The mixture of range runoff and lake water migrated downward through the fault-zones beneath the alluvial fans which served as conduits for fluid flow. On the basis of both isotope and ion geochemistry, the vertical component appears to dominate deep subsurface flow; there is no chemical or isotopic evidence for extensive interbasin flow in deep geothermal systems.

3. The Pleistocene lakes were not as large, widespread or deep in eastern Nevada. In these areas, it appears that the bulk of the recharge was supplied by runoff from the adjacent mountains directly to the fault conduits.

4. Alluvial fans represent the most likely point of fluid infiltration, from both runoff and rising Pleistocene lakes. Recharge during the Pleistocene was limited to a relatively narrow zone, below the level of the permanent ice pack as defined by nivation landforms, and above the impermeable fine-grained sediments that occupy playas and basin bottoms.

5. Most of the basins in Nevada and Utah are bounded on both sides by more than one set of faults. In the model presented here, (Figure 8.5) the fault zone nearest to the mountain range provides the principal conduit for recharge, and may also account for the system pressure differential that initiates fluid flow. Fault zones that are located closer to the basin act as recharge and discharge conduits, though not at the same location. This is consistent with the single fault plane model discussed by Welch (1981). The lateral flow model, also discussed by Welch (1981), is not directly supported by these data but a variation that does not require cross-basin flow may be postulated. In a lystric fault system (Figure 2.2), faults on one side of the basin are believed to taper basin-ward and eventually merge in a single structure. It is therefore possible to speculate that a lystric fault system completes the structural circuit required for the model shown in Figure 8.5, by connecting the recharging fault to the discharging

Figure 8.5: New conceptual model of geothermal recharge involving high-angle Basin and Range style faulting, range run-off, and fluid contribution from

large pluvial lakes that saturated the alluvial fans above faults. See text for discussion of recharge/discharge flowpaths.

fault. Minimum depths of circulation within the fault zones vary from 3 to 5 km and flow rates (from surface to depth to surface) range from 0.1 to 0.2 m per year.

6. In summary, geothermal fluids that currently discharge in the Great Basin are of Pleistocene origin. Figure 8.5 summarizes the principal components of the proposed model, showing the relationships between rising Pleistocene lake levels, runoff, alluvial fans, recharging and discharging fluids, and range-bounding faults. It should be noted that Figure 8.5 represents a two-dimensional view of the conceptual model. Recharging and discharging fluids, depicted by arrows, may occur within the same fault zone, but are well separated in space and possibly time. The degree of participation by Pleistocene lakes is generally proportional to lake size and depth, and basin geometry. Geothermal systems were recharged locally by Pleistocene-age precipitation, that in some cases mixed with rising lake waters which had flooded adjacent alluvial fans and underlying fault zones. Thermal characteristics are a complex function of depth of penetration of principal fault zones and proximity of recent volcanic activity.

CHAPTER 9. SUMMARY AND CONCLUSIONS

In the last decade, geothermal energy resources have evolved from a geologic curiosity to a major energy development industry in the Great Basin. During this evolution, a variety of resource assessment techniques have been applied in an attempt to enhance our understanding of geothermal and hydrothermal systems. Investigations designed to characterize geothermal systems have also evolved from basic geochemical studies to complex projects that integrate scientific disciplines outside of the realm of geothermal energy. This report describes the results of an interdisciplinary investigation and the development of a new model that explains the origin of geothermal fluids in the Great Basin.

The project consisted of three phases: literature search and preliminary data base compilation, field sampling and analysis, and data integration and interpretation. Several inter-related data sets were generated from various sources, including non-thermal and thermal fluid ion and isotope geochemistry, and paleoclimatic studies including paleohydrology and glaciology, and paleoclimate proxies.

Geothermal fluids throughout the world are known to be related to local precipitation, but in many cases are isotopically lighter. Regional analyses of stable isotopes of geothermal fluids in the Great Basin show that they are more depleted in heavier stable light-isotopes (lighter) than modern precipitation. In addition, the analyses illustrated a systematic variation across the study

area that was nearly identical to the distribution of the non-thermal meteoric proxy.

Paleoclimatological data from ice core in Greenland and deep-sea sediment core in the sub-polar Indian Ocean showed that the isotopic content of precipitation is not constant over time. Deuterium vs. time curves for ice core illustrated that a major depletion, relative to modern values, lasted from 30,000 to 10,000 years BP throughout the world. The deuterium content of wood from packrat middens from sites in the Great basin was analyzed and found to vary in a manner similar to the core data. Radiocarbon dated thermal fluids analyzed for deuterium showed a similar relationship to the paleoclimate data.

Paleohydrologic data showed that the levels of the largest lakes that occupied the Great Basin during the glacially dominated late Pleistocene (Bonneville, Lahontan, Russell, and Searles) varied systematically over time in response to changes in the climate. The lake level variation was also found to be nearly synchronous with the deuterium depletion record. The location of the lakes was found to be related to anomalous areas of both chemical and isotopic enrichment.

An interpretation of the data sets indicated that geothermal fluids that discharge at the surface (including those from pumped wells) originated as precipitation during the late Pleistocene, 10,000 to 30,000 years BP. The principal recharge sites are porous alluvial fans, adjacent to lake-filled basins, but below the snow and ice fields that occupied the Great Basin.

A model is proposed that challenges the previously held view that modern precipitation is the recharging fluid for geothermal waters that discharge from hot springs. Depths of penetration of circulating fluids are estimated to be 3 to 5 kilometers at an average rate of 0.1 to 0.2 m per year.

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APPENDIX 1. LABORATORY METHODS

This section provides an inventory of the various laboratories utilized and a brief description of the analytic procedures that were followed during execution of this investigation. References are provided for those wishing a more thorough description of certain procedures.

HYDROGEN AND OXYGEN STABLE LIGHT ISOTOPES

Water samples were collected from springs and wells in a manner to avoid contamination from either the atmosphere or stagnant, fractionating water which would result in enrichment of the sample.

Spring fluids were collected with a one liter polyethylene bottle attached to a retractable three meter sampling device, from as close to the spring's upwelling or issuing source as possible. The fluid samples were transferred to 200 ml glass bottles and the lids were immediately sealed with melted wax to ensure isolation.

Geothermal wells issuing above 100°C must be sampled with a mini-cyclone steam separator to avoid flashing (boiling) and partial loss of sample. The cyclone provides separate samples of the brine and steam phases, as well as the brine to steam ratio in the flow. Isotope analyses for the separate phases are recombined to provide the actual isotopic ratios, using:

$$(\% \text{ steam})(\delta \text{ steam}) + (\% \text{ brine})(\delta \text{ brine}) = \text{actual } \delta. \quad (\text{A1.1})$$

For example, a Beowawe, Nevada, power plant fluid sample containing the following phase fraction and deuterium concentrations:

	Fraction	$\delta D^{\circ}/_{\text{OO}}$
Brine	.8576	-130.5
Steam	.1424	-138.3,

recombines to give:

$$(.1424)(-138.3^{\circ}/_{\text{OO}}) + (.8576)(-130.5^{\circ}/_{\text{OO}}) = -131.4^{\circ}/_{\text{OO}}. \quad (\text{A1.2})$$

The technique for combining other chemical data, such as oxygen isotopes or geochemistry, is the same.

Isotope analysis consists of converting the sample to a gas (CO_2 for oxygen-18, H_2 for deuterium), which is then ionized and passed through a mass spectrometer for counting. Detection limits are $\pm 2^{\circ}/_{\text{OO}}$ for deuterium and $\pm 0.2^{\circ}/_{\text{OO}}$ for oxygen-18. Stable light-isotope analyses were contracted with the Environmental Isotope Laboratory of the University of Arizona, Tucson. Analysis of water samples GF-1 through GF-39 were completed for Oxygen-18 by the University of Arizona, however an equipment failure prevented completion of the contract. Deuterium analysis for samples GF-1 through GF-39 and complete analyses for samples GF-40 through GF-48 was performed at the University of Waterloo, Ontario, Canada.

WATER CHEMISTRY ANALYSIS

Water chemistry samples were collected in one liter plastic bottles and submitted to the University of Utah Research Institute (UURI) for analysis. Table A1.1 lists the techniques used for analysis of the elements included in the data set.

Table A1.1 - Water chemistry analysis techniques used by UURI.

METHOD	ELEMENT(S)
Inductively Coupled Plasma	Ca, Na, K, SiO ₂ , Mg, Li, B
Laboratory Titration	HCO ₃ , Cl
Specific Ion Electrode	F
Gravimetric	TDS
Turbidimetric	SO ₄

TRITIUM ANALYSIS

Tritium samples were collected from selected springs and deep wells in one liter glass bottles and sealed with Teflon tape. Tritium analysis was performed by the Desert Research Institute Analytical Laboratory of the University of Nevada System, Reno, Nevada. Samples were distilled by hydrolysis, followed by slow electrolysis at a constant temperature to evaporate the hydrogen (protium and deuterium) and concentrate the tritium. Counting was then performed with a scintillation counter. The process requires three months to complete and has detection limits of 5 TU.

CARBON ISOTOPES

Carbon isotope analysis was performed by Beta Analytic, Coral Gables, Florida. Approximately 4 grams of carbon is required for each analysis. Two types of material were submitted for carbon isotope analysis, carbonate precipitated from thermal waters and carbonate scale collected from geothermal production wells. The analysis reported ¹⁴C activity and the ¹³C/¹²C ratio.

Water Samples

Thirty-five gallons of water for each sample were collected from selected hot springs and geothermal power plant production wells deemed to have sufficient carbon for analysis. Wells that use chemicals to inhibit carbonate scale were not sampled.

The water was placed in a stainless steel carbonate precipitator and titrated with 50 to 500 ml sodium hydroxide to raise the pH to approximately 10. One liter of 8 molar strontium chloride solution was then added to produce a strontium carbonate precipitate according to the reaction:

The speed at which the precipitate formed and settled out varied from instantaneous to several days. Some of the samples produced a gelatinous precipitate which proved, upon analysis, to be devoid of carbon. A more detailed description of the carbonate precipitator and procedure can be found in Szecsody (1982).

Carbonate Scale

Calcium carbonate scale removed from geothermal power plant production wells was also submitted for analysis. Surface contamination was mechanically removed from the samples prior to submission. The samples were etched with acid by Beta Analytic in the laboratory to further remove any surface contamination.

Analytical Method

The water samples are filtered to remove the strontium carbonate, which is allowed to dry. Samples (water and scale) are then acidified to generate CO₂ gas which is purified in a vacuum. The sample is then synthesized to benzene. Stable isotopes, ¹³C and ¹²C, are counted in a mass spectrometer while unstable ¹⁴C is measured by a proportional counter. Detailed description of these procedures and equipment can be found in Faure (1986) and Fritz and Fontes (1980).

PACKRAT MIDDENS

Packrat middens were collected by Dr. Peter Wigand of the Quaternary Sciences Center, Desert Research Institute, Reno, Nevada. Initial treatment involves separation of the constituent macrofossils from the midden followed by sorting in a vertical blower.

Hydrogen and oxygen isotopes

All of the packrat midden hydrogen and oxygen analyses from this study were completed at the University of Arizona and the, Department of Biology, University of Miami, Miami, Florida. Juniper twigs from 25 middens in western Nevada were obtained from Dr. Wigand and submitted for analysis.

Midden analysis requires liberation of hydrogen and oxygen isotopes from the alpha cellulose structure of plant macrofossils

removed from the midden. An excellent account of the preparation and analysis of cellulose for stable light-isotopes is available in Siegel (1983).

Carbon-age determination

The middens had been previously dated for Dr. Wigand at the Beta Analytic laboratory. Fecal pellets from the same stratigraphic horizon as the twigs provided the carbon for dating. Duplicate dates on pellets have yielded ages agreeing within 2%, well within the analytical error.

APPENDIX 2. THEORY OF ENVIRONMENTAL ISOTOPES

The atomic structure of most matter consists of a nucleus, containing protons and neutrons, surrounded by electrons. The number of protons in an atom dictates the atomic number and the elemental properties of that atom.

Isotopes, different forms of an element, possess the same number of protons but a variable number of neutrons. Since isotopes have the same number of protons they occupy the same location in the periodic table of the elements, and derive their name from the Latin roots iso = same, and topos = position (Aswathanarayana, 1985). The combined total of protons and neutrons determines the atomic mass (measured as atomic mass units or AMU) of an element. A convenient way to denote isotopes is to place the atomic mass to the upper left and the atomic number to the lower left of the elemental symbol. Oxygen contains 8 protons and either 8, 9 or 10 neutrons, hence $^{16}_8\text{O}$, $^{17}_8\text{O}$, and $^{18}_8\text{O}$. In practice, only the atomic mass, not the atomic number, is normally displayed. The atomic number is implied by the elemental symbol.

Approximately 1,700 isotopes have been identified, of which only 256 are stable (non-radioactive). Most elements are a mixture of two or more isotopes; only 21 elements are purely composed of a single isotope. In multiple isotope elements, one isotope is usually predominant, the others present in only trace amounts (Hoefs, 1987).

Unstable (radioactive) isotopes "decay" to a different isotope or element by emitting atomic particles. They will be discussed later. Stable isotopes do not undergo decay, but may be exchanged with other materials during chemical equilibrium reactions.

STABLE LIGHT-ISOTOPES

Isotopes of "lighter" elements (less than 24 AMU) have been shown to be effective natural indicators of environmental conditions (Fritz and Fontes 1980, Faure 1986). The large percent difference between the atomic masses of isotopes of light elements allows the relative abundance of each isotope to be easily quantified through instrumental analysis, such as mass spectrometry. For example, the mass difference of hydrogen (1 AMU) and deuterium (2 AMU) is calculated by:

$$\frac{{}^2\text{H}-{}^1\text{H}}{{}^1\text{H}} = \frac{2 - 1}{1} = 1 = 100\%, \quad (\text{A2.1})$$

while the difference in oxygen, isotopic mass of 16 AMU and 18 AMU, is calculated by:

$$\frac{{}^{18}\text{O}-{}^{16}\text{O}}{{}^{16}\text{O}} = \frac{18 - 16}{16} = .125 = 12.5\%. \quad (\text{A2.2})$$

These percent differences are far more significant than that of the heavier elements. For example, uranium has isotopic masses of 238 AMU and 235 AMU:

$$\frac{{}^{238}\text{U}-{}^{235}\text{U}}{{}^{235}\text{U}} = \frac{238 - 235}{235} = .0128 = 1.28\%. \quad (\text{A2.3})$$

The stable light-isotopes of three elements, hydrogen, oxygen and carbon, are employed in this study. Isotope ratios are usually measured as deviation from an internationally accepted standard, normally assigned the value 0. They are reported in units of permil (parts per thousand), and indicated by the symbol ‰. Samples of the accepted standards are distributed to laboratories for instrument calibration to assure reproducibility of results. The standards that are used vary among elements and will be discussed individually below.

Hydrogen

The element hydrogen contains one proton and has three isotopes, two stable and one unstable. They are shown in Table A2.1 , along with their relative worldwide abundance. The internationally accepted standard for measurement of hydrogen stable isotope ratios in water is known as SMOW (Standard Mean Ocean Water). It was initially standardized by Craig (1961). A sample of water with the same composition as SMOW, referred to as Vienna-SMOW, is maintained by the International Atomic Energy Agency (IAEA) in Vienna, Austria and distributed to laboratories worldwide for equipment calibration. The deuterium(D)/hydrogen(¹H) ratio present in this water is assigned a value of ‰, and all other values are reported as a deviation (δ) from this standard, using the formula:

$$\delta D^{\text{‰}} = 10^3 * \frac{(D/^1H)_{\text{SAMPLE}} - (D/^1H)_{\text{SMOW}}}{(D/^1H)_{\text{SMOW}}} . \quad (\text{A2.4})$$

Table A2.1: Isotopes employed in this study and their average worldwide abundances from Aswathanarayana (1985).

ISOTOPE	COMMON SYMBOL	P	N	AMU	PERCENT ABUNDANCE	COMMENTS
Hydrogen	H, ¹ H	1	0	1	99.984	unstable
Deuterium	D, ² H	1	1	2	0.0154	
Tritium	T, ³ H	1	2	3	≤10 ⁻¹⁵	
Oxygen-16	¹⁶ O	8	8	16	99.763	not used
Oxygen-17	¹⁷ O	8	9	17	0.0372	
Oxygen-18	¹⁸ O	8	10	18	0.1995	
Carbon-12	¹² C	6	6	12	98.90	unstable
Carbon-13	¹³ C	6	7	13	1.10	
Carbon-14	¹⁴ C	6	8	14	≈10 ⁻¹⁰	

P - # of protons in the element
N - # of neutrons in the isotope

AMU - Atomic Mass Units

SLAP, a second standard distributed by the IAEA, is based on the D/H content of Standard Light Antarctic Precipitation and has a value of -428⁰/₀₀ on the SMOW Scale. All values in this study are reported relative to Vienna SMOW.

On a relative scale, more positive permil values contain more deuterium and are considered to be enriched in heavy isotopes while the more negative values are considered to be depleted in heavy isotopes (Figure A2.1). Enriched samples are commonly referred to as "heavy", while depleted samples are considered to be "light".

Figure A2.1: Deuterium isotope relative depletion/enrichment scale.

Oxygen

The element oxygen contains eight protons and has three stable isotopes, as shown in Table A2.1, along with their relative world-wide abundance. The accepted standard for oxygen isotope ratios in water is also Vienna SMOW, the $^{18}\text{O}/^{16}\text{O}$ ratio found in SMOW is zero permil. Oxygen-17 exists in such minute quantities it is not used for isotope studies. The formula for oxygen permil deviation from SMOW is similar to that for hydrogen:

$$\delta^{18}\text{O}^0/\text{‰} = 10^3 * \frac{(^{18}\text{O}/^{16}\text{O})_{\text{SAMPLE}} - (^{18}\text{O}/^{16}\text{O})_{\text{SMOW}}}{(^{18}\text{O}/^{16}\text{O})_{\text{SMOW}}}. \quad (\text{A2.5})$$

The IAEA also distributes a SLAP standard for oxygen, corresponding to $-55.5^0/\text{‰}$ on the SMOW scale. A different standard is used to express the isotope composition of oxygen in carbonates, and is discussed in the carbonate section that follows.

Carbon

As shown in Table A2.1, the element carbon contains six protons and has two stable isotopes. The internationally accepted standard for carbon isotopes is the Pee Dee Belemnite (PDB). The PDB standard ($0^0/\text{‰ PDB}$) corresponds to the $^{13}\text{C}/^{12}\text{C}$ ratio of carbon dioxide (CO_2) gas prepared from fossil Belemnites collected from the Cretaceous Pee Dee Formation in South Carolina, USA. Deviation from the PDB standard is calculated in the same manner as deviation from SMOW for hydrogen and oxygen isotopes:

$$\delta^{13}\text{C}^0/\text{‰ PDB} = 10^3 * \frac{(^{13}\text{C}/^{12}\text{C})_{\text{SAMP}} - (^{13}\text{C}/^{12}\text{C})_{\text{PDB}}}{(^{13}\text{C}/^{12}\text{C})_{\text{PDB}}}. \quad (\text{A2.6})$$

A PDB standard also exists for the oxygen isotopes of carbon compounds wherein the $^{18}\text{O}/^{16}\text{O}$ of the CO_2 gas liberated is assigned the value $0^0/_{\infty \text{ PDB}}$. The relationship between the PDB and SMOW oxygen standards is:

$$\delta^{18}\text{O}^0/_{\infty \text{ SMOW}} = 1.03037 * \delta^{18}\text{O}^0/_{\infty \text{ PDB}} + 30.86^0/_{\infty}. \quad (\text{A2.7})$$

All of the original PDB standard sample has been consumed for analysis, but several other carbon standards are in use today for reference and calibration. They include the National Bureau of Standards (NBS)-16 and -17 carbon dioxide gas ($-41.48^0/_{\infty \text{ PDB}}$ and $-4.41^0/_{\infty \text{ PDB}}$, respectively), NBS-20 Solenhofen Limestone ($-1.06^0/_{\infty \text{ PDB}}$), and the NBS-21 Graphite ($-28.10^0/_{\infty \text{ PDB}}$) (Hoefs, 1987).

ISOTOPE FRACTIONATION PROCESSES

The partitioning of isotopes between two substances or phases is known as isotope fractionation. For environmental isotopes, isotope exchange is the primary process of concern. Variation in temperature (a function of latitude and elevation) and degree of "rain-out" (distance from vapor source) are the primary controlling factors affecting exchange. Other processes include kinetic effects, diffusion, filtration, chemical composition and crystal structure (Hoefs, 1987).

Isotope Exchange

Isotope exchange reactions are similar to general chemical equilibrium and can be represented by:

with the equilibrium constant, K , for the reaction being:

$$K = \frac{(A_2/A_1)^a}{(B_2/B_1)^b}. \quad (A2.9)$$

Introducing the concept of the partition function, Q , demonstrates the reliance of fractionation on temperature (Hoefs, 1987). The calculation to determine Q for any species is:

$$Q = \sum_i (g_i e^{-E_i/kT}), \quad (A2.10)$$

and the equilibrium constant becomes:

$$K = \frac{(Q_{A2}/Q_{A1})^a}{(Q_{B2}/Q_{B1})^b}. \quad (A2.11)$$

Derivation of the Q calculation is beyond the scope and focus of this project. The important factor to note is the strong dependence upon temperature, T , due to its position in the exponential function.

Of more interest than the equilibrium constant discussed above is the isotope fractionation factor, α . It is defined as the ratio of the isotopic ratio (R_X) in two compounds or phases (A & B):

$$\alpha_{(A,B)} = \frac{R_A}{R_B}, \quad (A2.12)$$

where $\alpha = K^{1/n}$, K is the equilibrium constant and n is equal to the number of atoms exchanged (Hoefs, 1987). For example, the frac-

tionation factor for the exchange of ^{18}O and ^{16}O between water and calcium carbonate (CaCO_3) according to the equilibrium equation:

is given by:

$$\alpha = \frac{(^{18}\text{O}/^{16}\text{O})_{\text{CaCO}_3}}{(^{18}\text{O}/^{16}\text{O})_{\text{H}_2\text{O}}} = 1.031 \text{ at } 25^\circ\text{C}. \quad (\text{A2.14})$$

Since temperature is a critical factor in the determination of K , it is equally critical for α as displayed graphically in Figure A2.2.

As just demonstrated, a close relationship exists between temperature and isotopic ratios. At lower temperatures (higher latitudes and/or higher elevations), water vapor is isotopically

Figure A2.2: Fractionation factor for deuterium and oxygen-18 (Fritz and Fontes, 1980).

lighter (more depleted). The relationship between isotope content of precipitation and temperature was determined to be linear by Dansgaard (1964) as:

$$\delta^{18}\text{O} = 0.695T - 13.6, \quad (\text{A2.15})$$

and subsequently by Yurtsever (1975) as:

$$\delta^{18}\text{O} = (0.521 \pm 0.014)T - (14.96 \pm 0.21). \quad (\text{A2.16})$$

Gat (1981) demonstrated that the relationship is not linear, but actually is best represented by a series of regional curves converging near $\delta^{18}\text{O} = 0\text{‰}$ and $T = 30^\circ\text{C}$. This temperature relationship is shown graphically in Figure A2.3 for different types of recording locations worldwide..

Atomic mass is a crucial factor affecting many of the thermodynamic properties (i.e. vapor pressure, boiling and freezing point, fractionation, etc.) of an element. Heavier isotopes possess a reduced vibrational frequency between atoms and have a correspondingly higher inertia than the lighter isotopes of an element. A critical property in the fractionation of water vapor is vapor pressure, which is directly proportional to mass. In a given mixture, the more active, lighter molecules are preferentially evaporated, leaving behind the heavier molecules, a process known as Rayleigh distillation. The resulting vapor is therefore depleted in the heavier isotopes while the remaining liquid is enriched in the heavier isotopes. Figure A2.4 can be used to determine the Oxygen-18 value of water vapor or condensate at any given temperature.

Figure A2.3: Depletion of ^{18}O permil values with decreasing temperature for stations worldwide (Gat, 1981). Dashed lines on D) represent best

fit estimates of Dansgaard and Yurtsever. Solid lines are second degree polynomial best fit lines estimating Gat's converging regional curves.

The isotopes of hydrogen and oxygen combine in water nine different ways, forming molecules with masses that differ by as much as 20%. The possible combinations, along with their masses and percent variation from the predominant form (H_2^{16}O), are listed in Table A2.2. An example of how these mass differences affect some of the thermodynamic properties of water is shown in Table A2.3.

As just discussed, vapor leaving a body of water will be relatively depleted (lighter) in the heavier isotopes (that is, it will display more negative permil values), in comparison to the original source. During formation of clouds from this vapor, the heaviest molecules evaporated will be more likely to condense. The rain drops formed are relatively isotopically enriched and closely

Figure A2.4: Effect of temperature and remaining original vapor on ^{18}O concentrations. The process is the cause of rain-out and is known as Rayleigh distillation. The condensate line represents the isotopic composition of precipitation from the water vapor.

Table A2.2 - Possible combinations of hydrogen and oxygen isotopes in water and percent variation compared to H_2^{16}O .

FORM	AMU	%	FORM	AMU	%	FORM	AMU	%
H_2^{16}O	18	--	HD^{16}O	19	06	D_2^{16}O	20	11
H_2^{17}O	19	06	HD^{17}O	20	11	D_2^{17}O	21	17
H_2^{18}O	20	11	HD^{18}O	21	17	D_2^{18}O	22	22

resemble the composition of the original water source. As the cloud moves inland, the heavier molecules continue to be preferentially condensed and precipitated ("rain-out"). The isotopic content of vapor in the cloud, and subsequent precipitation from the cloud, becomes continually more depleted in the heavy isotopes (Figure A2.5).

Fractionation continues to take place after the droplet falls from the cloud. Isotopic exchange now occurs between the droplet and water vapor in the surrounding air. As in the initial evaporation process, the lighter molecules are preferentially evaporated,

Table A2.3 - Affect of hydrogen isotopes on the thermodynamic properties of water (Hoefs, 1987).

PROPERTY	CONDITION	H_2^{16}O	D_2^{16}O
Mass (AMU)	---	18.0	20.0
Density (g cm^{-3})	@ 20°C	0.9979	1.1051
Melting Temp. (°C)	@ 760 Torr*	0.00	3.81
Boiling Temp. (°C)	@ 760 Torr	100.00	101.42
Vapor Pressure (Torr)	@ 100°C	760.00	721.60

* 1 Torr = 1 mmHg, 760 Torr = 1 Atmosphere

causing the rain drops to become enriched in the heavier stable light-isotopes. Length of fall is therefore important in determining the final isotopic ratios within the precipitation. Precipitation that reaches lower elevations undergoes a longer fall and a longer period of exchange, and becomes more enriched isotopically than precipitation falling at higher elevations. Duration of each rainfall event is also important. The atmosphere will approach isotopic equilibrium with the precipitation, slowing or halting the fractionation process. Initial precipitation from a storm is therefore isotopically more enriched than subsequent rainfall from the same storm.

Figure A2.5: Diagrammatic example of storm rain-out. As a cloud moves inland, the vapor is progressively more depleted in the heavier isotopes (deuterium and oxygen-18) as they are preferentially condensed and removed in precipitation.

Other Fractionation Processes

Several other processes can effect isotope fractionation, but none to the extent of exchange. Some of the processes are briefly mentioned here, as obtained from Hoefs (1987).

- Kinetic isotope effects occur when the rate of chemical reaction is mass sensitive at a particular position. These effects can give clues to reaction pathways.
- Diffusion is primarily of importance for oxygen isotopes in rock-water interaction at elevated temperatures. The process is very complex and varies between different minerals.
- Bonds to ions with high ionic potential and low atomic mass have a tendency to preferentially incorporate heavier isotopes making chemical composition important.
- Heavier isotopes are likely to be more concentrated in more closely packed or ordered crystal structures.
- Ultrafiltration occurs during passage of water through semipermeable membranes such as clays and shales. Heavier isotopes are preferentially adsorbed at the clay minerals, shifting the residual waters retained by the clays to more positive permil values. Water that passes through the clay membrane becomes isotopically depleted (Coplen and Hanshaw, 1973). Laboratory experiments have demonstrated a deuterium depletion of $2.5^0/_{\infty}$ and an oxygen-18 depletion of $0.8^0/_{\infty}$.

UNSTABLE LIGHT-ISOTOPES AND AGE DATING

Most isotopes are unstable and decay spontaneously in a fission reaction, at a constant rate, to produce a daughter element and energy. The standard exponential decay reaction is:

$$N = N_0 e^{-\epsilon t} \quad \begin{array}{ll} N & = \text{current amount} \\ N_0 & = \text{initial amount} \\ \epsilon & = \text{decay constant} \end{array} \quad \begin{array}{ll} t & = \text{time} \\ e & = \ln^{-1} 1. \end{array} \quad (\text{A2.17})$$

Decay constants are virtually unaffected by temperature and pressure (Aswathanarayana, 1985). A common reference point for a radioactive substance is its "half-life" ($T_{1/2}$), which is the amount of time required for one-half of the substance mass to decay. By rearranging reaction A2.17 to:

$$t = - \frac{\ln(N/N_0)}{\epsilon}, \quad (\text{A2.18})$$

and setting the ratio N/N_0 equal to one-half, the half-life of an unstable isotope can be calculated from:

$$T_{1/2} = 0.693/\epsilon. \quad (\text{A2.19})$$

The two unstable isotopes included in this study, tritium and ^{14}C , both decay by emission of a high-energy, negatively charged beta particle (β^-). In this process a neutron disintegrates to a proton and a β^- , raising the atomic number by one producing a different element. β^- particles are similar to electrons in mass and charge, differing only in the mode of formation.

Tritium

Tritium (^3H), the unstable isotope of hydrogen, is naturally produced in the upper atmosphere by cosmic ray neutron (n) bombardment of nitrogen-14 according to the reaction:

The natural production rate is approximately 0.5 ± 0.3 atoms $^3\text{H}/\text{cm}^2/\text{sec}$ (Faure, 1986). Tritium concentrations are measured in tritium units (TU). One TU is equal to one tritium atom per 10^{18} atoms of hydrogen, and is equivalent to a tritium activity (A) of 7.1 disintegrations per minute per liter of water (dpm/l) (Faure, 1986). A disintegration is the spontaneous decay of one atom. Natural background concentrations of tritium are normally less than 25 TU (Fritz and Fontes, 1980), but locally vary greatly with latitude, altitude and season. The advent of above-ground nuclear weapons testing in the 1950's flooded the atmosphere with man-made tritium, increasing the background levels by two to three orders of magnitude. With the cessation of above-ground testing in 1963, atmospheric tritium levels have almost returned to pre-testing background levels (Faure, 1986). Minor quantities of man-made tritium continue to enter the atmosphere from the nuclear and defense industries.

Tritium, with a half life of 12.26 years, decays to helium by emission of a β^- particle by:

After isolation from the atmosphere, the decay of tritium occurs at a constant rate according to the standard exponential decay equation (A2.17). The short half-life of tritium limits its usefulness as an indicator of groundwater age to less than 50 years (Campana, 1981). It is nearly impossible to obtain meaningful quantitative dates from tritium samples due to the extreme difficulty in quantifying the initial activity (A_0) of the sample. Most researchers have found it useful to qualitatively associate high tritium activity of groundwater with precipitation that fell in the late 1950's and early 1960's.

Carbon

Carbon has one radioactive isotope, ^{14}C , which is also produced in the upper atmosphere by cosmic ray neutron (n) bombardment of nitrogen (^{14}N). The reaction involves replacing a proton in the nitrogen nucleus with a neutron:

The ^{14}C is rapidly oxidized to $^{14}\text{CO}_2$ and mixes with the atmospheric CO_2 reservoir, becoming available for incorporation into the hydrosphere and biosphere.

Certain types of material exposed to the atmosphere (living matter, water, etc.) undergo constant isotopic exchange with the atmosphere (Figure A2.6) maintaining an equilibrium ^{14}C activity of $A_0 = 13.56$ disintegrations per minute per gram (dpm/g). This standard activity is 95% of the activity of NBS-4990 oxalic acid in

1950 (Fritz and Fontes, 1980). Activity of ^{14}C is frequently reported as percent modern carbon (PMC), and is calculated by:

$$A_{\text{PMC}} = \frac{(^{14}\text{C}/^{12}\text{C})_{\text{SAMP}}}{(^{14}\text{C}/^{12}\text{C})_{\text{STAND}}} * 100, \text{ STAND} = 13.56\text{dpm/g.} \quad (\text{A2.23})$$

A sample with a PMC of 100 is equivalent to 13.56dpm/g.

When water is isolated from the atmosphere, or when organisms die, isotopic exchange ceases and the ^{14}C activity (A) decreases as the unstable ^{14}C decays. The decay process is a beta decay and the reaction is:

Figure A2.6: The carbon cycle, and the concentration of carbon expected in various materials that may contribute 'dead' carbon to groundwater.

The "uncorrected" age of the sample is obtained by rearranging the standard exponential decay equation (A2.17) to give:

$$t = \frac{-T_{1/2}}{\ln 2} \ln \frac{A}{A_0} \quad \begin{array}{l} A = \text{measured sample activity} \\ t = \text{age} \quad T_{1/2} = 5730y \\ A_0 \approx 13.56\text{dpm/g.} \end{array} \quad (\text{A2.25})$$

All carbon-age dates are calculated as radiocarbon years prior to a reference date of 1950.

CORRECTION OF CARBON AGE DATES

Carbon-age calculation assumes that the atmospheric activity (A_0) of ^{14}C has been constant throughout time and that no exchange of carbon has occurred with the surroundings since the sample was isolated and ceased isotopic exchange with the atmosphere. Both assumptions are invalid and correction factors must be incorporated to account for these effects.

Correction theory involves accounting for the amount of "dead" carbon introduced after exchange with the atmosphere has ceased. Activity of ^{14}C is measured as the number of disintegrations per minute that occur per gram of total carbon present. If additional carbon introduced to the sample after isolation provided no initial ^{14}C activity ("dead" carbon), the measured ^{14}C activity will be attributed to a greater initial volume of carbon than was actually present. The calculated age is then older than the actual age. Referring to Figure A2.7, the ^{13}C content of several types of material likely to be encountered and exchange isotopes with groundwater during circulation is demonstrated.

Atmospheric Correction

Seuss (1955) showed that A_0 of ^{14}C from 20th century wood was 2% lower than that of 19th century wood (Suess effect), due to the combustion of fossil fuels and the corresponding introduction of "dead" carbon to the atmosphere (Figure A2.8). Subsequently, DeVries (1958) showed that A_0 of ^{14}C from 1500 to 1700 AD was 2% greater than in the 19th century (DeVries effect). The reason for this change is unclear. More recent studies indicate that A_0 regularly fluctuates, within limits of $\pm 5\%$ (Fritz and Fontes, 1980). A comparison between ^{14}C dates from tree rings to their dendro-chronological age (Michael and Ralph, 1970), has provided correction tables valid to 8000 years Before Present (BP). Correction tables for older dates were developed from seasonal glacial lake

Figure A2.7: Carbon-13 content of various materials that may interact with groundwater and introduce 'dead' carbon during circulation.

varves (Stuiver, 1970a). Above ground nuclear weapons testing in the late 1950's introduced large quantities of ^{14}C into the atmosphere resulting in a marked increase in the atmospheric A_0 of ^{14}C .

Soil and Rock Correction

The science of correcting carbon-age dates for interaction of the water with soil (organic input), rock (dissolution) and clay (cation exchange) is, unfortunately, not as advanced as the atmospheric corrections. To make a fully accurate correction, the conditions present at the time of recharge and the flow path the water has followed must be determined. This is a worthy but currently unrealistic goal.

Figure A2.8: The Suess effect. Depletion of atmospheric Carbon-14 values that has occurred over the past 150 years (Stuiver, 1981) due to the combustion of fossil fuels and the resulting introduction of 'dead' carbon into the atmosphere.

Carbon incorporated in the soil zone is derived primarily from root respiration, and the correction factor depends upon the vegetation type (Table A2.4). The factors listed have been determined through field measurement of soil gas carbon content.

The unquantifiable unknown in this correction is the soil condition at the time of fluid recharge, several thousand to tens of thousands of years BP. Many authors have used present day soil conditions as a proxy for paleo conditions (Szecsody 1982; Glancy 1986), but this approach is flawed due to the rapid rate of Pleistocene/Holocene climate fluctuations. Use of paleo climate indicators to suggest conditions at the time of recharge is a better approach, but is also flawed. Correction factors can result in adjustments of several thousand years, possibly invalidating the paleo condition assumption that was made to perform the initial correction. This type of circular reasoning does not provide a reliable correction.

Table A2.4: Correction factors for soil zone carbon contribution from vegetation (Szecsody, 1982).

PLANT TYPE	TYPE	CLIMATE	FACTOR
CALVIN PHOTOSYNTHETIC	C4	COLD REGION	-25
HATCH-SLACK PHOTOSYNTHETIC	C3	ARID REGION	-12
SUCCULENTS STRESSED	C4		-25
UNSTRESSED	C3		-12
NO VEGETATION, ONLY ATMOSPHERIC ¹³ C IN SOIL			-6.4

Proper correction for rock-water interaction is equally challenging. Precisely determining the flow path of the fluids, the rock types encountered and the physio-chemical conditions present at the time of encounter is impossible. Estimates can, however, be made based upon the rock types and temperatures likely to be encountered, and geochemical ratios present. Isotopic exchange with limestone, for example, is the most likely source of dead carbon input. The correction factor depends on whether the limestone is a freshwater ($-4.93 \pm 2.75^{\circ}/_{\infty \text{ PDB}}$) or saltwater ($+0.56 \pm 1.55^{\circ}/_{\infty \text{ PDB}}$) carbonate (Faure, 1986). Figure A2.7 shows the range of ^{13}C values for various rock types.

The activity for soil (sl) and rock (rk) can be calculated by:

$$^{14}\text{A}_{\text{corr}} = ^{14}\text{A}_{\text{measured}}/P, \text{ where } P = \text{sl}/(\text{sl}+\text{rk}), \quad (\text{A2.26})$$

and:

$$\delta\text{C}_{\text{samp}} = \frac{\text{sl}}{\text{sl}+\text{rk}} \delta\text{C}_{\text{sl}} + \frac{\text{rk}}{\text{sl}+\text{rk}} \delta\text{C}_{\text{rk}}, \quad (\text{A2.27})$$

which rearranges to:

$$P = \frac{\delta\text{C}_{\text{samp}} - \delta\text{C}_{\text{rk}}}{\delta\text{C}_{\text{sl}} - \delta\text{C}_{\text{rk}}}. \quad (\text{A2.28})$$

$\delta^{13}\text{C}/^{12}\text{C}$ permil values are then inserted to calculate the correction factor, P.

Cation exchange with clays relates to the ultrafiltration discussed previously. The heavier isotopes are preferentially adsorbed by the clay minerals isotopically depleting the water as it passes through. Any clayey soil or rock could serve to slightly isotopically deplete the groundwater flowing through it.